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Implications of Global Warming for Canadian East Coast Sea-Ice and Iceberg Regimes Over the Next 50 to 100 Years

ATMOSPHERIC ENVIRONMENT SERVICE

DOWNSVIEW, ONTARIO

1991

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IMPLICATIONS OF GLOBAL WARMING FOR CANADIAN EAST COAST SEA-ICE AND ICEBERG REGIMES OVER THE NEXT 50 TO 100 YEARS

by

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ABSTRACT

This report presents the results of a scoping study which was initiated in response to concerns that atmospheric and oceanic changes related to increased greenhouse gas emissions may have significant effects on the iceberg and sea-ice regimes off the east coast of Canada. The study concluded that sea ice extent was one of the key parameters explaining variability in the number of icebergs crossing south of 48°N. This connection was considered to be consistent with physical processes affecting iceberg transport and deterioration, namely, the role of sea ice in reducing iceberg deterioration rates by damping wave erosion and holding water temperatures below 0°C. Analysis of sea ice and iceberg severity data, and proxy indicators over the period 1880-1990 showed no convincing evidence for a systematic amelioration of ice conditions, except for a short period of rapid decline in ice severity during the 1960's. On the contrary, since about 1960, winter regional temperatures and ice severities were found to have been directly out-of-phase with global temperature trends.

General Circulation Models (GCM's) and empirical analogue techniques were evaluated for their utility in scenario construction. GCM results were not considered reliable for the study area because of inadequate representation of ocean currents, poor spatial resolution, and simplified parameterizations of sea-ice thermodynamics and neglect of ice advection. Investigation of warm-cold temperature and pressure analogues did not support the presence of a consistent regional response to larger-scale warming. The sensitivity of sea ice to year round regional warming was investigated through Bilello (1960) melting degree day relationships. The analysis revealed an approximate 1 degree latitude retreat in the southern ice boundary off eastern Canada for every degree of warming. The estimated ice extent change for 4 to 6°C annual warming was comparable to the ice extent difference noted between extreme high and low ice extent years over the past 20-30 years. An investigation of glacier calving response to climate warming from mass balance computations indicated that iceberg calving rates were likely to decrease under significant annual warming (+6°C). However, the calculation ignored transient glacier response which could generate short periods of higher calving rates.

On the basis of current knowledge, it was concluded that ice and iceberg severities were unlikely to undergo major changes in the immediate future. However, over the next 50-100 years, if greenhouse gas forcing did result in year round regional warming of 2 to 4°C, sea ice retreat was considered to be likely. Such a retreat would, *inter alia*, result in reduced iceberg severity on the Grand Banks via increased potential for iceberg deterioration. A large number of uncertainties and knowledge gaps were identified in the study which need to be addressed before more reliable scenarios of iceberg severity change can be developed. Of these, the main areas identified for further research were related to obtaining a greater understanding of the mechanisms responsible for regional sea ice extent variability, glacier calving and transient glacier response.

RÉSUMÉ

Ce rapport présente les résultats d'une étude de la portée des incidences entreprise à la suite de certaines préoccupations. On se demandait si les changements atmosphériques et océaniques liés à la hausse des émissions de gaz à effet de serre ne pourraient pas produire des effets importants sur le régime des icebergs et des glaces de mer au large de la côte est du Canada. Cette étude conclut que l'étendue de la glace de mer est un des principaux paramètres qui expliquent la variabilité du nombre des icebergs qui descendent au sud de 48°N. On a estimé que ce lien cadre avec les phénomènes physiques qui touchent le déplacement et la détérioration des icebergs, c'est-à-dire avec la rôle de la glace de mer dans la baisse du taux de détérioration des icebergs, glace qui atténue l'érosion par les vagues et qui maintient la température de la mer au-dessous de 0°C. L'analyse des données sur la glace de mer et l'ampleur des icebergs, ainsi que des indices indirects de la période 1880-1990, n'a révélé aucune preuve probante d'une amélioration systématique de l'état des glaces, si ce n'est pour une brève période de rapide déclin de l'ampleur des glaces pendant les années 1960. Au contraire, depuis 1960 environ, on a constaté que les températures régionales et que l'ampleur des glaces d'hiver étaient complètement déphasées par rapport aux tendances des températures du globe.

On a évalué l'utilité, pour la construction des scénarios, des modèles de circulation générale (MCG) et des techniques analogiques empiriques. Du fait d'une mauvaise représentation des courants océaniques, d'une mauvaise résolution spatiale, d'un établissement simplifié des paramètres de la thermodynamique de la glace de mer et de l'omission de l'advection des glaces, on a estimé que les résultats des MCG n'étaient pas fiables pour la zone étudiée. A l'étude des données analogiques des pressions et des basses et hautes températures, on n'a pas décelé de réaction régionale cohérente au réchauffement à plus grande échelle. D'après les rapports des degrés-jours de fonte de Bilello (1960), on a examiné la sensibilité de la glace de mer au réchauffement régional survenant à longueur d'année. Cette analyse a détecté, par degré de réchauffement, un recul d'environ 1 degré de latitude dans la limite sud des glaces au large de la côte est du Canada. La modification estimative de l'étendue des glaces pour un réchauffement annuel de 4 à 6°C est comparable à l'écart observé dans l'étendue des glaces entre les années de maximum absolu et de minimum absolu de l'étendue des glaces au cours des 20 à 30 dernières années. En étudiant le vêlage des glaciers face au réchauffement du globe d'après les calculs d'équilibre des masses, on a établi que le taux de vêlage des icebergs allait sans doute décroître sous l'effet d'un important réchauffement annuel (+6°C). Toutefois, ces calculs n'ont pas tenu compte de la réaction transitoire des glacier, qui pourrait engendrer de courtes périodes de taux plus élevé de vêlage.

D'après les connaissances actuelles, on a conclu qu'il n'est guère probable que l'ampleur des glaces et des icebergs subisse d'importants changements dans un avenir immédiat. Toutefois, au cours des 50 à 100 prochaines années, si le forçage des gaz a effet de serre entraîne un réchauffement régionale de 2 à 4°C à longueur d'année, on considère comme probable un recul des glaces de mer. Vu les possibilités accrues d'une détérioration des icebergs, un tel recul engendrerait notamment une réduction de l'ampleur des icebergs aux Grands Bancs. Dans cette étude, on a relevé dans les connaissances nombre d'incertitudes et de lacunes qu'il convient de supprimer avant de pouvoir concevoir des scénarios plus fiables de la modification de l'ampleur des icebergs. A ce sujet, les principaux domaines où l'on a déterminé qu'il fallait pousser la recherche touchent l'amélioration de la connaissance des mécanismes qui causent la variabilité régionale de l'ampleur des glaces de mer, le vêlage des glacier et la réaction transitoire des glaciers.

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1. INTRODUCTION

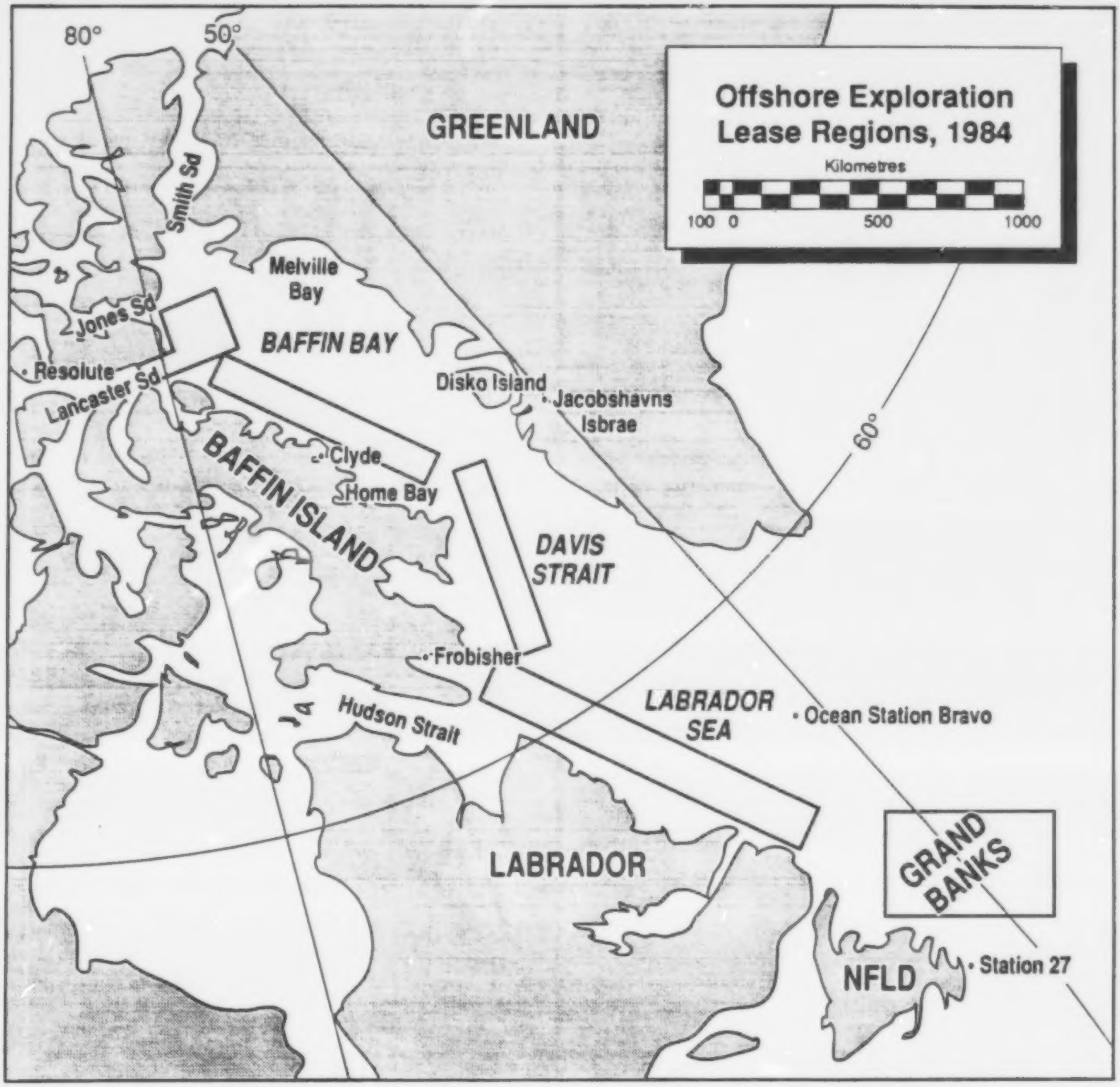
The severity of sea ice and iceberg concentrations off Canada's East Coast represents an important factor in determining the cost and feasibility of hydrocarbon-related development. Recently, concerns have been raised regarding possible impacts of global climate change on regional ice conditions (Roots, 1989). Interest is most immediate on the Grand Banks and Labrador Shelf, the locations of the two most southern and currently most active of the exploration lease areas delineated in Figure 1. The sensitivity of activities in these areas to the ice environment is illustrated by recent estimates (Stokoe, 1988) of additional exploration costs of as much as $30 million in years of extreme (severe) as opposed to average ice conditions. Consequently the suggested possibility (Roots, 1989) of substantial increases in the number of icebergs produced in regional source areas represents a potentially major impediment and hazard to a viable hydrocarbon development program, and, as well, poses significant problems for the marine transport and fishery industries.

The present study was initiated to examine the feasibility of providing meaningful assessments of changes in future ice/iceberg severities resulting from continuing anthropogenic modification of the atmosphere. In this study, the principal immediate effect of modification to the atmosphere will be assumed to be the overall global warming which is almost universally predicted by current global circulation models (GCMs) (Figure 2) (Wigley, 1989). The time period of concern extends from the present until late in the twenty-first century, when, according to the best current consensus, elevations of 2°C to 3°C may be expected in the annual global average atmospheric temperature.

A flowchart of the major environmental components involved in determining both present day sea ice/iceberg severities and their response to changed atmospheric $C0_2$ content is provided in Figure 3. The Figure depicts the important environmental systems, their key characteristic parameters and essential linkages and feedbacks. In spite of the implicit simplifications (Figure 3), the complexity of the connection between the severities and the changing $C0_2$ input to the atmosphere is evident. Consequently, efforts in this study will be, at least initially, concentrated on clarifying details of the processes represented in the upper-most and right-hand portions of the diagram which most directly affect the iceberg- and sea ice-severities. These severities will be assumed to be well described, respectively, by the number of icebergs observed south of 48°N during each spring-summer season (a parameter which has been monitored by the International Ice Patrol [IIP] since the early part of this century) and the spring sea ice areal extent off Labrador south of 55°N (the quantity of which has been the focus of recent regional studies (Ikeda, et al., 1988 and Peterson and Prinsenberg, 1990).

The prime area of interest for this study is the general locale of major hydrocarbon reserves off the Canadian East Coast, particularly in the Labrador Sea and Grand Banks regions. However, we also consider the source regions for ice and icebergs which extends to Davis Strait, Baffin Bay and associated side channels which may influence events in the downstream areas.

Our assessment begins in Section 2, with a brief review of the regional atmospheric, oceanic, sea ice and iceberg environmental components. The characteristics of the interannual variability of sea ice and iceberg conditions are summarized (Section 3) on the basis of both previous studies and some new approaches. Emphasis is given to identifying the sources and details of past and present variabilities. These identifications are then used to focus subsequent considerations (in Section 4) of climate change onto issues which are of most direct relevance to the ice/iceberg severities of concern. The latter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Melville Bay
Jones Sd
BAFFIN BAY
Resolute
Lancaster Sd
Disko Island
Jacobshavns Isbrae
Clyde
Home Bay
BAFFIN ISLAND
DAVIS STRAIT
Frobisher
LABRADOR SEA
Ocean Station Bravo
Hudson Strait
LABRADOR
GRAND BANKS
NFLD
Station 27

Figure 1 The greater east coast Canada region including offshore hydrocarbon exploration lease areas and geographical references.

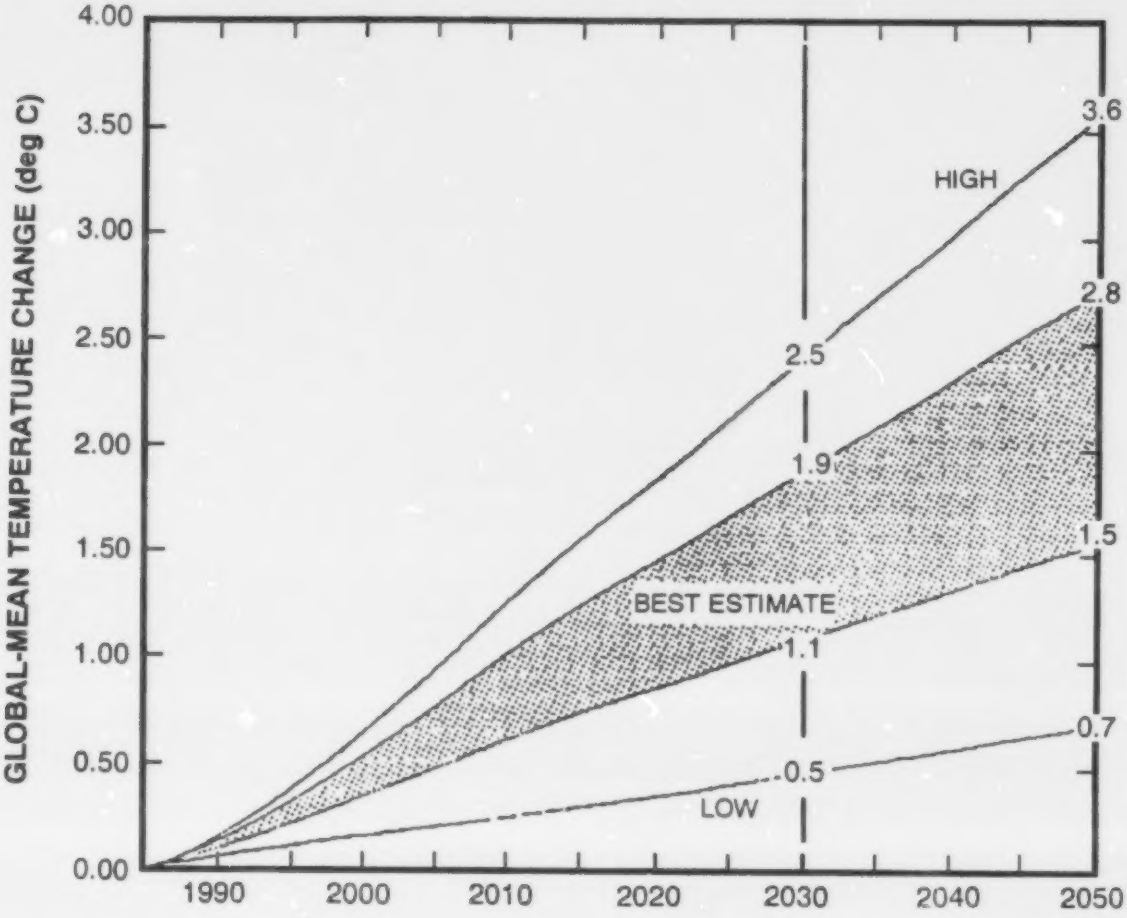

Figure 2 Projected change in annual global mean temperature based on present GCM model results (Wigley, 1989).

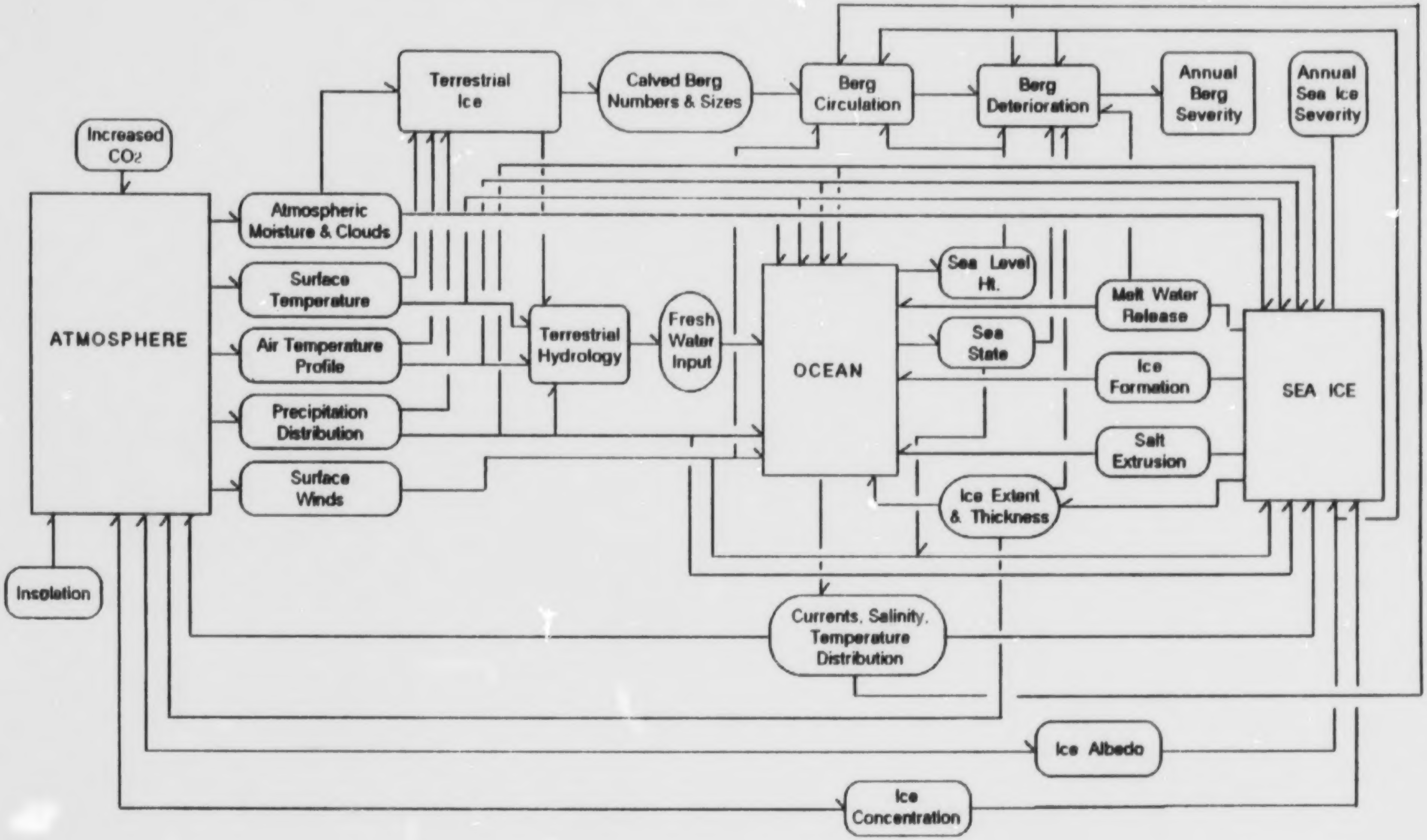
Increased CO2
ATMOSPHERE
Insolation
Atmospheric Moisture & Clouds
Surface Temperature
Air Temperature Profile
Precipitation Distribution
Surface Winds
Terrestrial Ice
Calved Berg Numbers & Sizes
Berg Circulation
Berg Deterioration
Annual Berg Severity
Annual Sea Ice Severity
Terrestrial Hydrology
Fresh Water Input
OCEAN
Sea Level Ht.
Sea State
Melt Water Release
Ice Formation
Salt Extrusion
Ice Extent & Thickness
SEA ICE
Currents, Salinity, Temperature Distribution
Ice Albedo
Ice Concentration

Figure 3 Flowchart representation of major environmental components, and their associated parameters and linkages.

considerations review alternative approaches to quantifying future climate change and evaluate their reliability and implications for the assessment requirement. Finally, in Section 5, a program for meeting this requirement is outlined and its feasibility assessed in terms of its necessary research components.

2. ENVIRONMENTAL BACKGROUND

2.1 The Atmospheric Environment

The climate within the entire study area varies considerably as would be expected from its significant latitudinal extent (45°-75°N). The main climatic influences are:

(1) solar radiation, which decreases by a factor of five or more from south to north;

(2) atmospheric momentum exchanges, driven by general wind patterns;

(3) the location of the entire study area being to the east (and generally downwind) of a large continental land mass, which profoundly affects air mass characteristics;

(4) rugged and mountainous terrain along much of the eastern coastline throughout the study area;

(5) the vast elevated ice cap of Greenland, glaciers and ice sheets on Ellesmere, Devon, Bylot and Baffin Islands, as well as the terrestrial snow cover which persists through most of the year in the more northerly portions of the study area;

(6) the presence of a highly variable sea ice cover.

Summaries and analyses of the general climate features of the study area have been conducted separately for Newfoundland/Labrador (Peach, 1975; Bursey et al., 1978; Petro-Canada, 1982; Phillips, 1990) and the Canadian Arctic (Maxwell, 1980; 1981). Of particular interest for the present study are storm activity, winds, pressure and air temperature distributions, since these quantities are the most important meteorological influences on sea ice and iceberg distributions in the study area.

The intensity and frequency of storms generally decreases from south to north, since most of the storms originate at lower latitudes. The most common winter storm tracks (Bursey et al., 1978; Brown et al., 1986; Maxwell, 1981) follow an east-northeasterly path over the northern Grand Banks and from N. Newfoundland to S. Labrador. Some of these winter storms, after moving into the Labrador Sea, track northerly and account for nearly all of the storm activity reaching Davis Strait and Baffin Bay.

Strong winds are associated with the storms passing through the region. The winds are highest during the autumn through winter to early spring period, when storm activity is more frequent than in the later spring and summer months. Marine winds measured in the central Labrador Sea at Ocean Weather Station Bravo (56.5°N 51.0°W) averaged 24.4 knots in January, compared to 11.9 knots in July. The frequency of winds exceeding 48 knots amounted to 3.4 percent in January, as compared to 0.1 percent in July. The most intense winds are associated with the passage of storms through the area. Typically, these strong winds occur over synoptic periods (a few to several days). They are usually associated with changing wind directions, often beginning with a southerly component, through either easterly or westerly directions (according to whether the storm centre passes to the south or north, respectively), followed by a northerly wind component. The storm activity often plays a moderating role in advecting relatively warm air into the region from the south, as noted by Maxwell (1981).

Another important influence in determining wind patterns is the role of the polar vortex associated with the vast area of high pressure situated over the Arctic Ocean throughout most of the year. In winter, this area of high pressure intensifies and expands southward to occupy the northern portions of Canada. In conjunction with the quasi-permanent Icelandic Low, usually centred between Greenland and Iceland, this results in a persistent pattern of westerly to northerly winds which dominate the longer-term atmospheric circulation of the study area. The extremely low temperatures advected from the north and the west play an important role in (1) formation of sea ice locally and (2) southward advection of sea ice from northerly portions of the study area to more southerly regions.

Because of the comparatively limited number of reliable and representative direct wind observations available over the ocean surface, atmospheric pressure distributions are generally used to derive the large scale wind patterns. The extended time series of pressure distributions available over mean years of observations can be used to examine the interannual variability of large-scale atmospheric circulation. Different variations to this approach have been applied in previous studies. One method is to use the computed North Atlantic Oscillation difference in atmospheric pressure between the mid-latitude (20° to 40° N) high pressure and the Icelandic Low pressure as an index of the general circulation patterns over the North Atlantic (Walker and Bliss, 1932). Another approach is to derive geostrophic winds at grid points over the North Atlantic from the pressure distributions, as in the maps of pressure and wind stress anomalies for 1950-1980 (Thompson and Hazen, 1983) or the time series of geostrophic winds for the period 1946 to 1988 as derived using the techniques developed by Swail (1985).

Air temperatures are of particular importance for this study. Temperatures represent a first order indicator of vertical heat exchanges between the atmosphere and ocean (or ice) surface. As in the case of winds, temperature measurements over the ocean surface are generally sparse. However, time series observations of extended duration are available from coastal stations, and unlike winds, coastal observations of air temperatures are, in the presence of high coastal and offshore ice uncertainties, reasonably representative of over-ocean or -ice conditions.

Extended time series of geostrophic winds and air temperatures in winter (Figure 4) for the northern Labrador and S. Baffin Island region suggest large variations occur over a wide range of time scales. Very low temperatures appear to occur at intervals of approximately 11 years, as indicated by anomalies exceeding -5°C in 1961, 1972 and 1983. For these same years, a rough correspondence to strong north to northwesterly winds can be discerned in the geostrophic wind data.

Low air temperatures and strong northerly winds result in more extensive sea ice cover off the Canadian east coast. A prominent example of very extensive sea ice coverage coinciding with cold temperatures and northerly winds occurred in 1972 (Saulesleja and Phillips, 1982). The association between large winter ice extent and cold northerly and westerly winds has been demonstrated by Crane (1978) for the Davis Strait/Labrador Sea region over the period 1964-1974. The interannual variability of ice, and its possible relation to atmospheric distributions, is considered further in Sections 3.2 and 4 of this study.

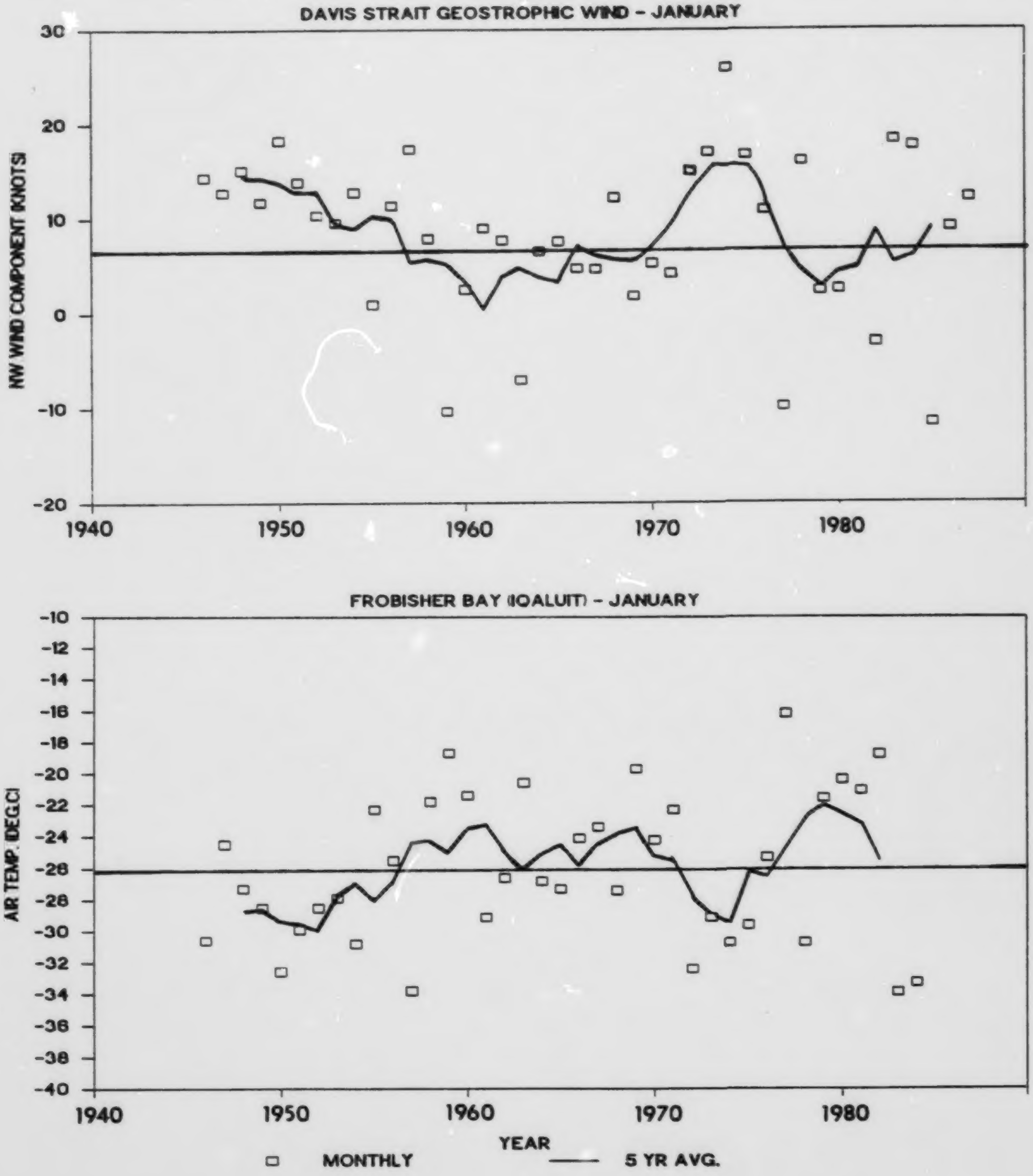

Figure 4 Temporal variability of January winds in the central Davis Strait area and corresponding mean January air temperatures at Frobisher Bay (Iqaluit).

Important investigations of atmospheric climatology have been conducted for subregions of the study area, primarily in the Eastern Canadian Arctic (Bradley, 1973; Barry et al., 1975; Barry and Kean, 1978; Crane, 1978; and Jacobs and Newell, 1979). The Baffin Island area is considered to be particularly sensitive to climatic fluctuations on both short and longer time scales (Barry et al., 1975). Bradley (1973) and Barry et al. (1975) noted that inter-decadal temperature trends differed considerably in the Baffin Island region from the mean Northern Hemisphere variations. They related the observed changes in temperature to differences in synoptic weather patterns. Crane (1978) showed that the differences for the summer season of ice retreat can be related to the relative displacement of the 700 mbar trough over the Eastern Canadian Arctic. In years for which this trough is displaced to the west (east), synoptic storm activity from the south is increased (decreased) resulting in warmer (colder) temperatures and early (late) ice retreat.

Recent results obtained by Walsh and Chapman (1990a), indicate strong negative correlations between pressures in the Arctic portion of our study region and in the mid-latitude Atlantic regions. These occur on monthly time scales and thus are not simply associated with climatological characteristics such as the tendency of Atlantic cyclones to penetrate the Arctic. This wider view of atmospheric forcing was taken a step further in later work (Walsh and Chapman, 1990b), which related major North Atlantic sea ice anomalies to atmospheric fluctuations over a large area extending from Greenland through the polar regions to northern Asia.

The North Atlantic Oscillation (NAO) has been, to some extent, correlated with interannual variability in almost all major environmental components off the east coast (Walsh et al., 1986; Myers et al., 1988). On global scales, the NAO may be teleconnected to other global oscillatory features, such as the El Nino Southern Oscillation (ENSO) phenomenon (Rogers and Van Loon, 1979; Namias, 1981). Similar relationships have been established between the sea level pressure and the 11 year solar-cycle during the westerly phase of the Quasi-Biennial Oscillation (Van Loon and Labitzke, 1988). According to Walsh and Chapman (1990), these relationships suggest a forced resonance in the atmosphere's internal dynamics which is excited by a variety of factors, including solar flux variations.

2.2 The Ocean Environment

The oceanic circulation in the study area is dominated by large-scale, buoyancy- and wind-driven currents found over the continental margins of the ocean basins. These currents (Figure 5) have been identified individually for specific margin segments as the West Greenland, Baffin and Labrador Currents. Recently, it has been hypothesized (Chapman and Beardsley, 1989) that these individual features are part of an interconnected flow system having a length of over 5,000 km., which represents the largest known coastal current in the world.

This large-scale current system (Chapman and Beardsley, 1989) extending from the southern tip of Greenland to the U. S. continental shelf, itself, may be one part of an even larger coupling of ocean circulation features. Some evidence for this coupling may be seen in subsequent effects arising from a reported (Dickson et al., 1988) event of widespread freshening of the 500 to 800 m layer of the North Atlantic Ocean. This phenomenon, the "Great Salinity Anomaly" (GSA) was first observed north of Iceland in the mid to late 1960's. Subsequently, similar freshening of the upper ocean was observed along the Labrador coast during 1971 to 1973, in the Norwegian Sea during 1977 to 1979, and most recently in the Greenland Sea during 1981-1982. Dickson et al. (1988) have explained these

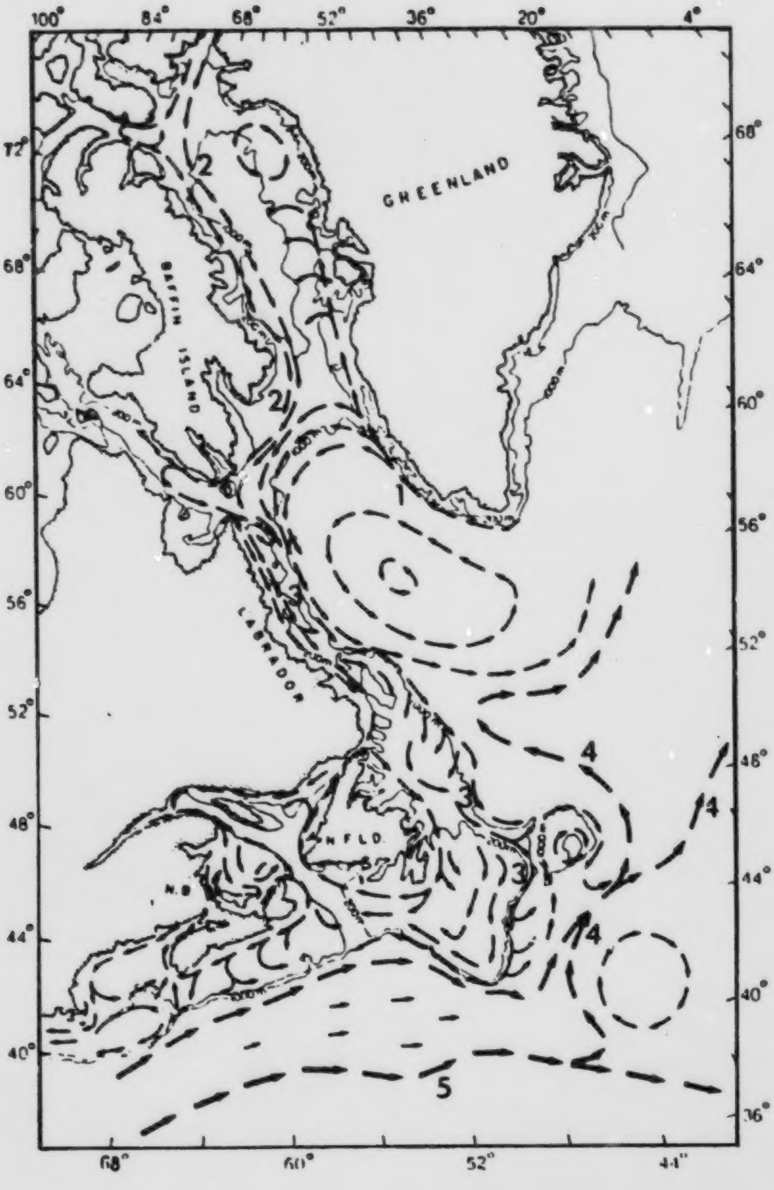
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Figure 5 Major surface current features of the study region (adapted from Trites, 1982). Individual currents shown are: 1) West Greenland Current; 2) Baffin Current; 3) Labrador Current; 4) North Atlantic Current; and 5) the Gulf Stream.

GSA observations as an advective event being transported around the northern portions of the North Atlantic Ocean by linked, large-scale circulation features.

The proposed linkage between our immediate area of interest off eastern North America and the Greenland, Iceland and Norwegian Seas (GIN seas) of the northeastern North Atlantic has important implications for the upper water column of our study area. In particular, major outflows of freshwater from the Arctic Ocean through the Greenland Sea (Aagaard and Carmack, 1989) may account for the GSA, which, through the postulated advection process, could have a significant impact on the Labrador Sea. Salinities which influence the degree of stratification in the upper ocean have potential global implications through inhibiting large scale heat exchanges between the atmosphere and ocean. A large degree of ventilation of heat from the deep ocean occurs in the Greenland Sea, with lesser occurrences in the Labrador Sea.

In general the features of the study region circulation can be classified into four main categories:

(1) narrow and often intense currents situated over the steep continental slopes; these strong currents follow the local bathymetry oriented with the coastline to the right of the flow direction. Typical and maximum velocities are 0.2-0.4 m/s and 0.7-1.0 m/s, respectively. This current type constitutes the main branches of the Labrador, Baffin and West Greenland Currents;

(2) weaker and more directionally variable currents which occur over the continental shelves and, in particular, over the extensive banks which constitute most of the shelf area;

(3) inshore currents of moderate strength which occur within 20 to 30 km of the coastline; associated typical and maximum velocities generally fall into ranges of 0.1-0.2 m/s and 0.35-0.5 m/s, respectively. This type of current is directed parallel to the slope current, although with reduced magnitude;

(4) large scale cyclonic gyres over the deep water basins of the Labrador Sea and Baffin Bay; the velocity magnitudes associated with these gyres are comparatively small (0.05-0.2 m/s) except under very strong local wind forcing.

The dominant forcing mechanisms for the continental slope currents [type (1)] are provided by the buoyancy fluxes resulting from geopotential differences associated with the presence of less dense waters over the shelf regions. The reduced density of waters over the shelf arise from glacial and land run-off, ice melt and the freshwater outflow of side channels such as Lancaster Sound and Hudson Strait (Smith and Schwing, 1990). Local winds, while important for upper layer currents in all areas, are the dominant forcing mechanism (Webster and Narayan, 1988) over the comparatively shallow banks of the shelf [type (2)] and in the inshore zone [type (3)]. However, in the latter region [type(3)], buoyancy fluxes from land run-off and ice melt can also be significant (Csanady, 1981). In the offshore regions [type (4)], large scale wind patterns result in the observed large scale cyclonic gyres (Thompson et al., 1986; Wright et al., 1986).

The intense slope currents are strongly influenced by the complex meanderings of the shelf edge found off the East Coast of Canada and West Greenland. Bathymetry is particularly important in the vicinity of adjoining water bodies (such as Smith, Jones and Lancaster Sounds in Baffin Bay and Hudson Strait in the northwest Labrador Sea) and in the many areas where underwater canyons traverse all or part of the width of the continental shelf. In such areas, the strong cores of the major currents, which are normally confined to the shelf edge, penetrate onto the shelf proper. The currents often describe re-entrant patterns into and out of the shelf (Fissel et al., 1982; Fissel and Birch, 1990; Peterson and Symonds, 1988; Petrie and Isenor, 1984).

Knowledge of the seasonal variability in the major regional ocean currents is derived from a small number of year-long records obtained in the study area. In the northernmost regions, the data available are limited to a few long-duration current meter records (e.g. Lemon and Fissel, 1982; Ross, 1990a, 1990b). Preliminary results from a nearly continuous current meter time series data set spanning ten years obtained off Labrador (Lazier, 1988) are also now available. These year-round data sets indicate that the offshore branch of the Baffin and Labrador Currents generally strengthens in autumn, attaining maximum flow speeds in September-November. Currents decrease through the winter months, with the smallest magnitudes occurring in April-May. Lazier (1988) notes that the seasonal change in the velocity of the offshore branches of the Labrador Current follows the same seasonal pattern as the difference in steric (internal) sea-level heights between the shelf and the abyssal regions of the Labrador Sea. The difference in steric heights from winter to summer is primarily associated with the seasonal cycle in salinity on the Labrador shelf, under the combined effects of land run-off and melting of sea-ice. The Labrador Current effectively confines the seasonal fresher water to the shelf regions.

The magnitude of interannual variability in the Labrador Current is also presently being studied. The ten-year (1978-1987) time series of current measurement in the Hamilton Bank region of the Labrador Shelf should provide much valuable information, but analysis results are not yet available. Myers et al. (1989) have used indirect estimates of volume transport over the upper 100 m of the water column, derived from water mass distributions, to examine interannual variations in the Labrador Current over a forty-year period at sites off southern Labrador, southeastern Newfoundland and West Greenland. Off southern Labrador (the Seal Island transect) the total transport during summer over the shelf and slope in the upper 100 m of the water column was estimated to vary by a factor of two among the studied individual years. The summer transports in southern Labrador, West Greenland and southeastern Newfoundland exhibited statistically significant negative correlations with the strength of the mid-latitude westerlies in the North Atlantic of the previous winter, as represented by the North Atlantic Oscillation Index (Myers et al., 1989). These results, although somewhat less certain due to the indirect means of estimating transports, do suggest that winter atmospheric and sea-ice conditions are important to interannual variability of the Labrador Current.

Water properties in the study area (see for example, Smith et al., 1937; Kiilerich, 1937; Dunbar, 1951) are notable for the comparatively low temperatures and salinities found in the upper portion of the water column (Figure 6) relative to those of the North Atlantic Ocean proper. These distinguishing water properties are most pronounced over the continental shelves on either side of the Labrador Sea, where large horizontal gradients occur over the continental slope separating the cold, fresh shelf waters and the warmer, more saline waters of the central deep water basins. The strong gradients in water properties, extending to depths of 500 m or more, mark the offshore branch of the major regional currents.

Within the upper layer of the water column over the shelf regions, water temperatures and salinities exhibit a very pronounced seasonal variation. Water temperatures exhibit a seasonal range from very close to the freezing point of seawater (-1.8°C) in winter to as

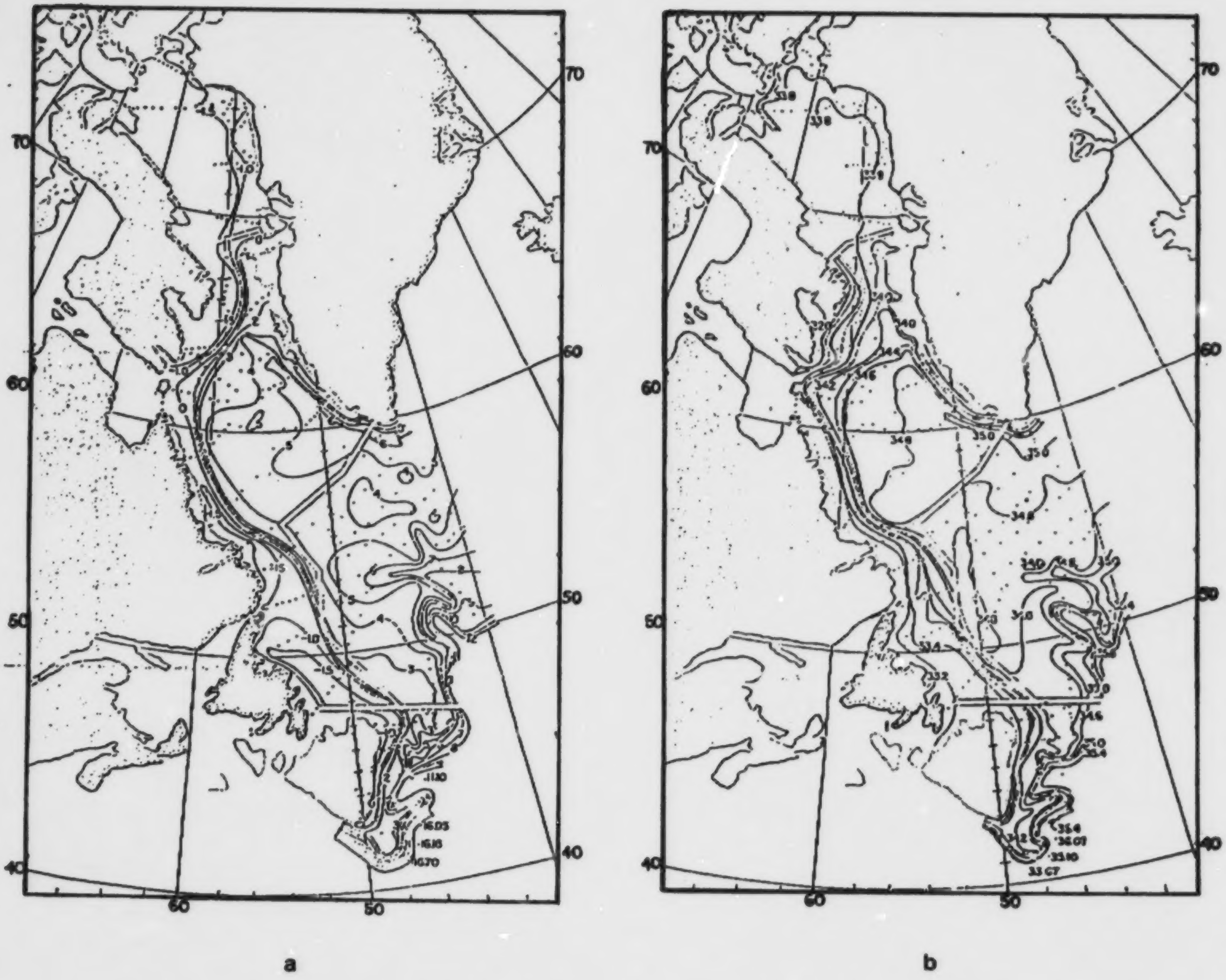


Figure 6 Distribution of temperature (a) and salinity (b) at 100m depth in the study region [summer-fall data] (Smith et al., 1937).

much as several degrees above this temperature during summer and autumn. The amplitude of the seasonal pattern increases at more southerly latitudes, as the duration of ice clearing increases. The seasonal range of surface temperatures is limited to 2-4°C in northern Baffin Bay (Fissel et al., 1982), but can reach 10°C off southern Labrador (Lazier, 1982) and even larger ranges of 15°C or more over the Newfoundland continental shelf (Keeley, 1981). The range of interannual variability in sea surface temperature is generally much smaller at 2 to 2.5°C, as estimated by Dunbar (1982).

Below the upper layer, cold (less than 0°C) temperatures persist over the shelf through the summer months and into autumn over most of the shelf width from Davis Strait to the Grand Banks. This water, termed the Cold Intermediate Layer (CIL), has a characteristic thickness of 100 to 200 m, and has minimum summer temperatures of -1.0° to -1.5°C (Petrie et al., 1988). The CIL is confined to the shelf proper, with temperatures increasing from the shelf edge offshore.

Interannual variations in the cross-sectional area and mean temperature of the summer CIL on the shelves of southern Labrador and Newfoundland appear to be spatially coherent (Petrie et al., 1988) and correlate well (r=.81 to .99 and -.86 to -.91, respectively) with the average ice extent south of 55°N from January to June. A case in point is the period 1983 to 1985, when sea-ice off South Labrador and Newfoundland extended unusually far to the south. During each summer of this period, the CIL was colder and larger than average. The unusually low temperatures of the CIL during 1983 to 1985, and a similar large negative temperature anomaly during 1972 to 1974, coincided with extreme meteorological and sea-ice conditions in the East Canadian Arctic.

2.3 Sea Ice

Information on sea ice in the study area has been collected in a reasonably systematic manner for the past 130 years off eastern Newfoundland (Hill and Jones, 1990). However, in northern areas, the mapping of sea ice extent only became operationally feasible with aircraft in the 1950's. In all areas, the advent of satellites for mapping large scale sea ice extent greatly improved data quality, from the mid-1960's to the present. Measurements of other sea ice properties such as ice types and ice thicknesses are much less extensive, both in terms of total duration of observations and in the spatial and temporal resolution of the data sets.

Extension of sea ice through the region begins in September in northwestern and western Baffin Bay with growth around a core of remnant summer ice which is exported into the region by the side channels of Lancaster, Jones and Smith Sounds (Marko, 1981; NOC, 1986). These channels are the major source of multi-year ice in the predominantly first year pack ice characteristic of the region. The combination of earlier freezing dates at higher latitudes and the cyclonic air and ocean current circulations advance the ice edge to the south with a pronounced lag in eastern- as opposed to western-Baffin Bay. Typically, large amounts of ice begin to appear in Davis Strait in November and off northern Labrador approximately one month later (Figure 7a).

By November large fields of fast (immobile) ice have stabilized extending from the shoreline to as much as 70 km offshore over the shallow continental shelves off eastern Baffin Island and off west Greenland. With the steady thickening of the ice cover (to eventual undistorted thicknesses of up to 2 m) sharp gradients in ice drift velocity tend to disappear and in Baffin Bay and Davis Strait 9+ tenths concentrations of ice, consisting mostly of vast floes, move relatively uniformly to the south at an average speed of

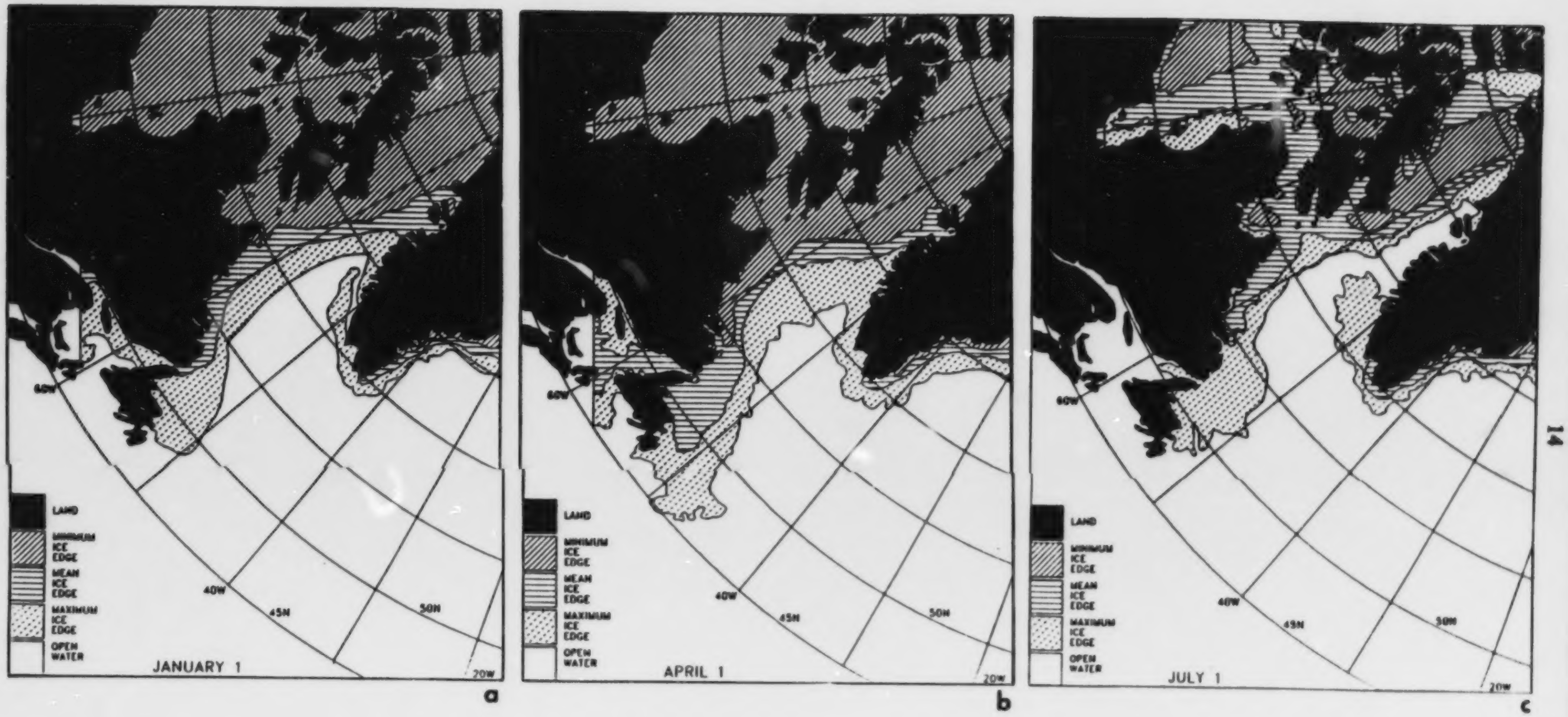

LAND
MINIMUM ICE EDGE
MEAN ICE EDGE
MAXIMUM ICE EDGE
OPEN WATER
60W
40W
45N
50N
20W
JANUARY 1
APRIL 1
JULY 1
a
b
c

Figure 7 Climatological mean, maximum and minimum sea ice extent off the Canadian East Coast as at (a) January 1, (b) April 1 and (c) July 1 [NOC, 1986].

approximately 15 km/day (Jordan and Neu, 1981). Off Labrador, the much thinner ice (.2 to .5 m), generally consisting of floes with diameters of 1 to 10 m, allows ice greater mobility over the entire continental shelf region with maximum velocities above the narrow outer branch of the Labrador Current centred on the continental slope.

The largest seasonal extent of the southern ice edge is usually achieved in March (Figure 7b). On average, substantial reductions from the maximum annual ice extents off Labrador and on the Grand Banks do not begin to be observed until April. The maximum offshore extent of the ice in this southern area rarely extends beyond the continental slope region. The trend in the retreat of the regional ice cover is a reversal of the advance pattern with the ice edge moving northward and westward most rapidly adjacent to Greenland as the western regions continue to receive large amounts of ice discharged by the side channels of Lancaster, Smith and Jones Sounds and from Hudson Strait (Figure 7c).

The interannual variability in sea ice extent is large and can be comparable to the climatological mean ice extent itself in southern regions (Figures 7a-c). This aspect of sea ice, which is central to the topic of this study, is specifically addressed in Section 3.

2.4 Icebergs

Introduction

The quantity of icebergs which reach the Grand Banks and northeast Newfoundland shelf each spring and summer depend on many environmental factors. To provide an overview of the complex atmospheric, oceanic and sea ice parameters involved, it is convenient to consider the iceberg environment off the Canadian East Coast in terms of (a) the supply of icebergs, as calved from glaciers; (b) the transport of icebergs from the source regions to the impact area of interest to this study; and (c) the mechanisms involved in the deterioration and eventual disappearance of icebergs.

Iceberg data sets are highly variable in measurement quality and their spatial and temporal extents. Counts of icebergs are the best measured iceberg parameter and data of this type have been recorded for well over 80 years in southern Newfoundland waters but have been confined to intermittent observations over perhaps 40 years in Baffin Bay. Even the annual iceberg counts obtained from extended observations in southern waters are prone to considerable uncertainties. This particular data set has been collected largely through the efforts of the International Ice Patrol of the U.S. Coast Guard and the Atmospheric Environment Service of Environment Canada. Uncertainty in annual iceberg counts arises in part due to the major changes in technology applied to detect icebergs, evolving from ship-based visual sightings to ship-based radar sightings and later to side-looking airborne radar (SLAR). Information on other iceberg parameters, including mass, shape, velocity, calving and grounding events, and glacial dynamics is much less extensive than the iceberg count data.

Iceberg densities over much longer periods, extending over thousands of years, can be estimated from geological measurements of iceberg scours on the seafloor (Lewis et al., 1987). However, large knowledge gaps, particularly in the rates of scour degradation, introduced large uncertainties and greatly limit temporal resolution for the interpretation of these data. Consequently, such results will not be considered further in this report.

Iceberg Supply

Icebergs in the region are predominantly generated at the tidal glacier fronts of West Greenland with smaller contributions (estimated at 10%) from East Greenland and the Arctic Archipelago. Annual production of iceberg mass from West Greenland, itself, has been estimated to be 200 km^3 (Reeh, 1985). Based on an estimated median iceberg mass of 0.5×10^6 tonnes in Baffin Bay (Marko, 1981) the total annual production off West Greenland would be equivalent to approximately 400,000 icebergs.

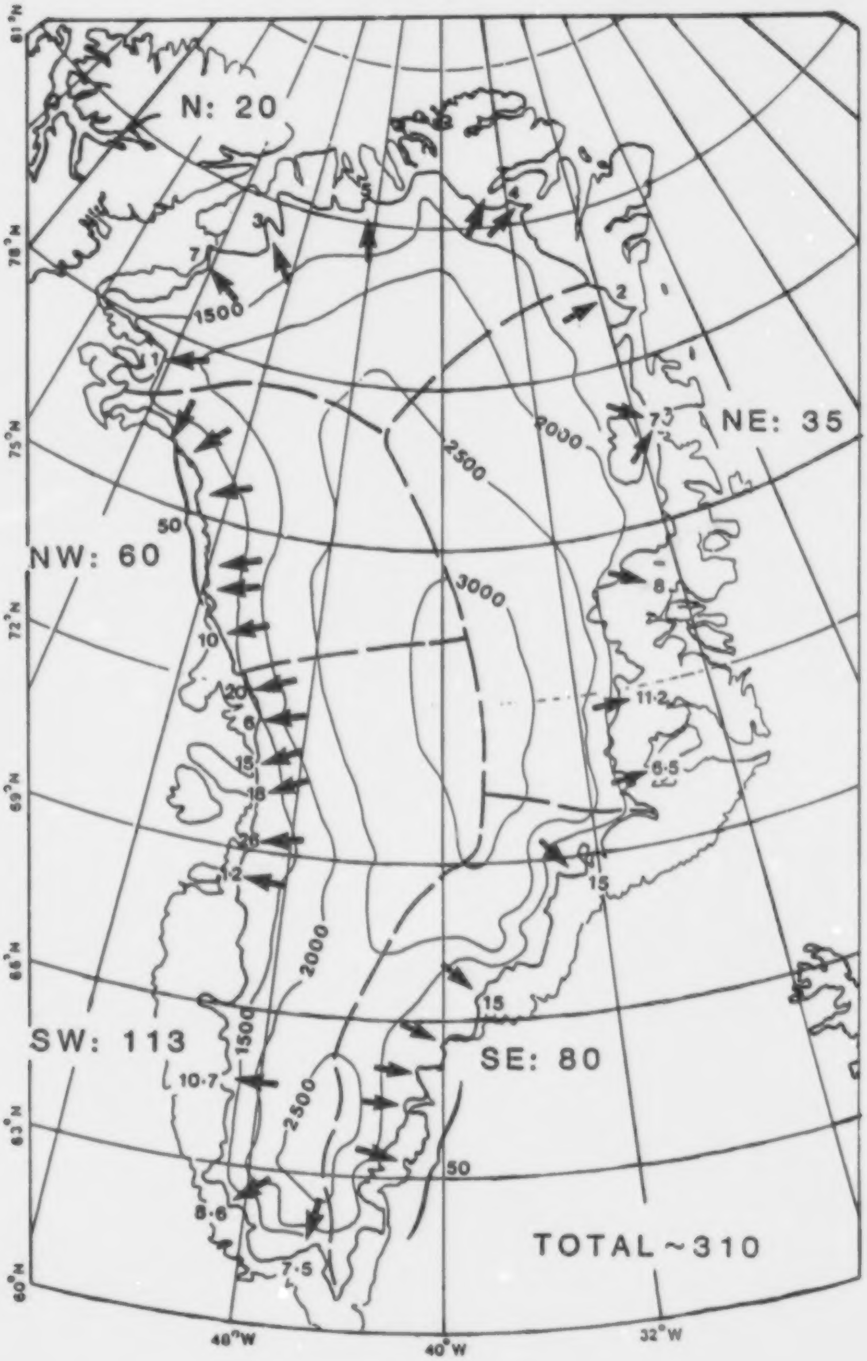
Recent estimates of the annual production of Greenland iceberg mass volume (in amounts of km^3/yr) show an approximate correspondence with adjacent levels of precipitation (Figure 8). Large values occur in the south, up the west coast as far as about 69°N and up the east coast as far as about 71°N. There is another pocket of high precipitation in the coastal region north of Melville Bay, at 76-77°N. Clearly these coastal regions of high precipitation correspond to regions where there is open water offshore for much of the year, with the Melville Bay anomaly being a result of the presence of the North Water winter polynya (Ito, 1982). This allows high evaporation and increased regional snowfall. The contrast with coastlines which are completely icebound in winter is striking. For example, on the west coast the variation in annual precipitation (water equivalent) is from about 20 cm in the icebound 70-75°N zone, to 50-60 cm from Disko Island (approximately 70°N) southward and 80-90 cm near Cape Farewell (approximately 60°N).

Iceberg Transport

Once calved, bergs in the region typically require a year or more (Birnie and Williams, 1985) to move away from the local areas adjoining glacier termini. In part, this delay is associated with the necessity for traversing the narrow, twisting fjords which separate many glacier fronts from the open sea. In most cases, however, delays are also associated with the presence of fast ice, both in these fjords and over shallower portions of the continental shelves. The demonstrated restriction of independent berg movement by even moderately thick ice (Canpolar, 1985) precludes significant iceberg escapement prior to fast ice break-up (Dowdeswell, 1989).

Once located in open water or in mobile sea ice over the continental shelves, bergs tend to follow trajectories which closely resemble the regional currents (Figure 5). The pattern of flow is again cyclonic with larger berg movements of 25 to 30 km/day centred in regions over the continental slope (at moderate or low ice concentrations). Slower more erratic movements prevail in other areas. Bergs tend to follow bottom contours producing distinctive cyclonic re-entrant loops above the numerous submarine canyons which penetrate the continental shelves. Over broader canyons such as that extending from the entrance to Lancaster Sound, the primary effect of these diversions is to lengthen trajectories (Figure 9). Icebergs moving along the steeper, narrower penetrations of the continental shelves are often brought close to areas of shallow water into which they can be diverted and grounded under the influence of winds and local currents. Particularly prominent regions where such diversions and groundings take place are located near Clyde and Home Bay off eastern Baffin Island and in Melville Bay and near Disko Island off West Greenland (Figure 1).

The importance of the diversion process was indicated by the trajectories of 21 bergs, fixed with satellite-trackable beacons (Marko, 1981; Marko et al., 1982) in the late summer months of 1978 and 1979 near the entrance to Lancaster Sound (Figure 9). Only five of the monitored bergs were observed to reach Davis Strait and none of these bergs crossed the


N: 20
NE: 35
NW: 60
SW: 113
SE: 80
TOTAL ~ 310
1500
2000
2500
3000
2000
2500
1500
2000
7
3
5
4
2
1
73
8
11·2
6·5
15
15
50
50
10
20
6
15
18
26
1·2
10·7
8·6
7·5
81°N
78°N
75°N
72°N
69°N
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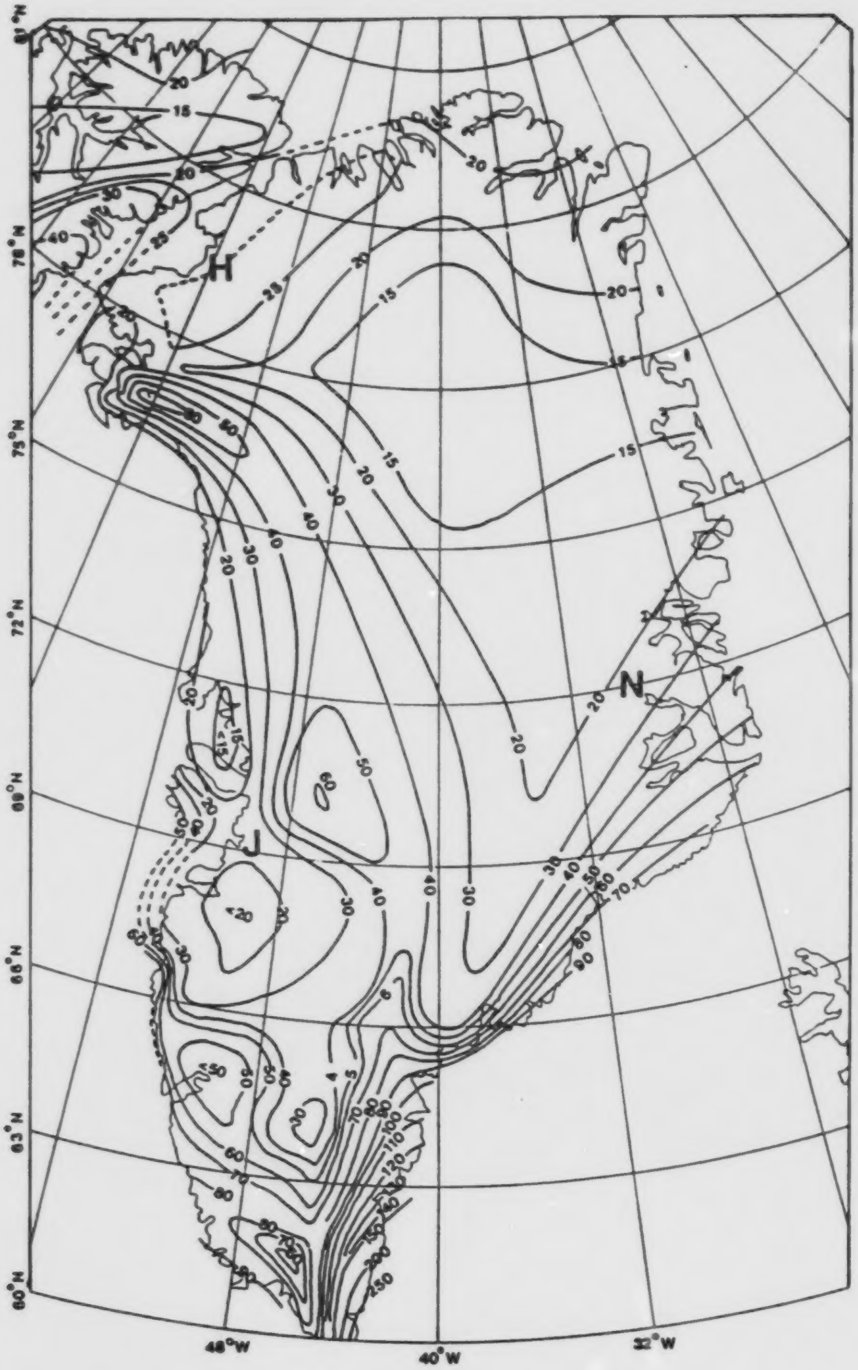
63°N
60°N
48°W
40°W
32°W

Figure 8a Map of estimated iceberg mass production in km^3/yr of water equivalent (Reeh, 1985).


81°N
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6

Figure 8b Map of present day precipitation in $g/cm^2/yr$ for Greenland (Reeh, 1985).

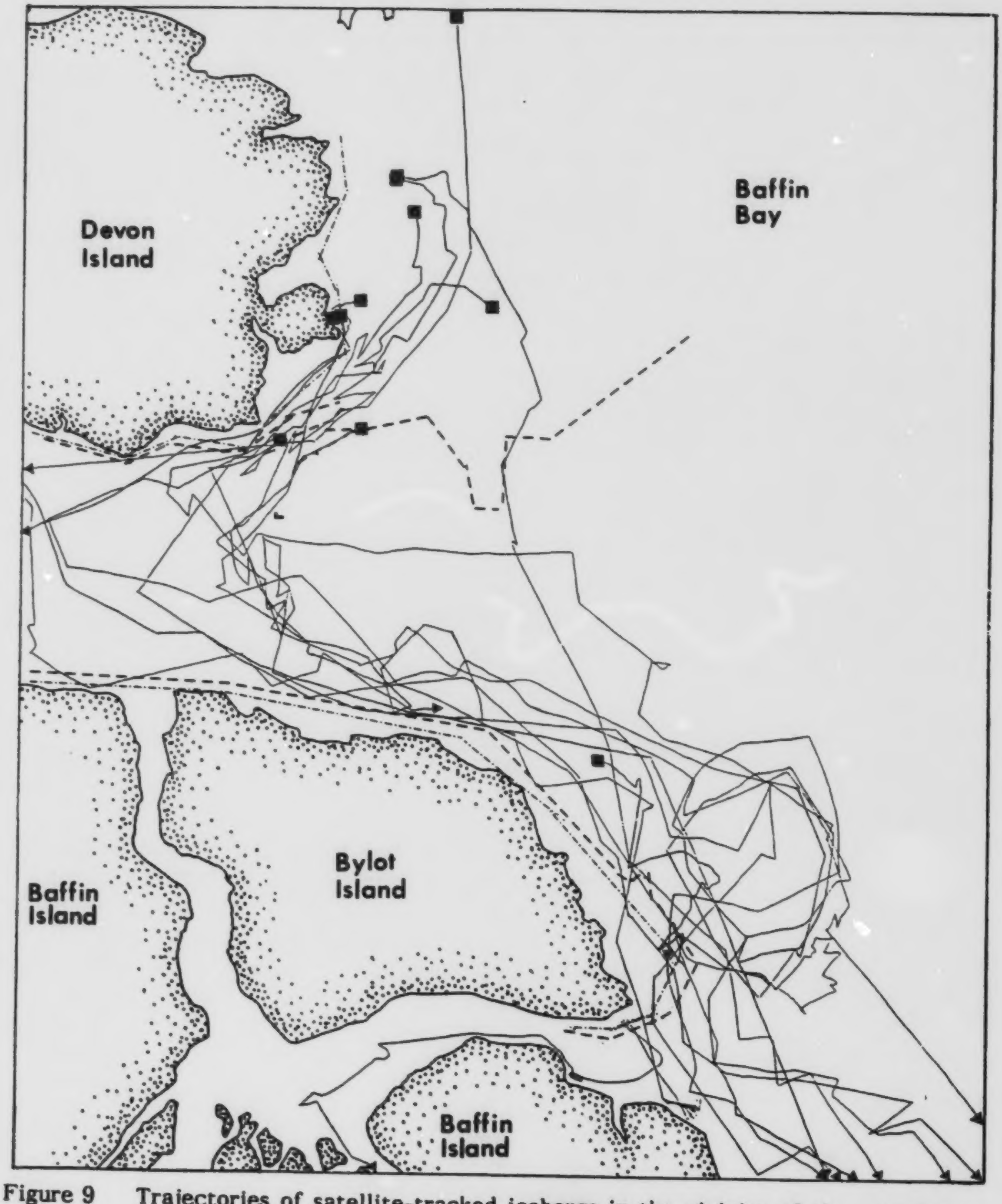

Devon Island
Baffin Bay
Baffin Island
Bylot Island
Baffin Island

Figure 9 Trajectories of satellite-tracked icebergs in the vicinity of the entrance to Lancaster Sound (Marko, 1981). The solid squares indicate the berg locations at the (August-September, 1978) start of tracking. Broken lines denote the approximate 200 and 500 m bathymetric contours.

67°N northern boundary of the Strait in less than eight months. The discrepancy between these travel times and expectations (100 days), assuming even a minimal 10 km/day southerly drift, was due to repeated grounding of the icebergs. The bergs remained grounded for substantial time periods, ranging up to 15 months and beyond.

One of the 21 tracked icebergs reached Labrador within one year. In that instance, the extended track of the berg began in November, 1980. The already extensive fast ice present at that time over the continental shelf areas would have precluded grounding during the winter months. Similar occurrences of unimpeded (by grounding/entrapment) southerly drift are believed to be responsible for the sudden increase which usually occurs in the Davis Strait berg population in the January-February period. This increase coincided with the approximate arrival times (assuming the 15 km/day average drift rate of the regional mobile sea ice) of those bergs which have moved past the upstream Baffin Island grounding areas in the post October period associated with the presence of the fast ice buffer. Similar mechanisms of seasonal ice modulation of berg fluxes can be assumed to affect the northerly berg flow adjacent to Greenland.

The above key features of the berg circulation lead to the identification of at least three distinct components in the movement of bergs toward the Labrador and Grand Bank regions of interest. These components are:

(1) berg fluxes associated with the strong, relatively narrow southerly continental slope current which moves maximum numbers of bergs into Davis Strait during the January-July period, when the upstream coastal shelves of west Baffin Bay are shielded by fast ice;

(2) berg fluxes which move southward through the abyssal offshore waters of Baffin Bay following the westward and then southward diversion of the West Greenland Current between 73°N and 75°N;

(3) berg fluxes associated with the similarly westward and then southward diversion of the southern portion of the West Greenland Current which occurs between roughly 65°N and 67°N at the northern, constricted end of Davis Strait.

Unfortunately, at present, we have little direct data on the relative strengths of these individual flux components. With respect to the first 2 components, estimates were made previously on the basis of late summer and early fall surveys of Baffin Bay in 1948, 1949, 1968 and 1969 (Morgan, 1971). The data of 1948 suggested that roughly two thirds of the observed bergs were contained within the first flow component. A slightly smaller proportion of icebergs (55%) was observed in the component 1 flow in 1969, while in 1949, 69% of the bergs were reported to be associated with the central Baffin Bay flow component 2.

Merging and convergence of all three berg flow components would appear to occur in northern Davis Strait, producing high concentrations over the strong current, continental slope region. Bergs moving through this region during the months of July-November are subject to high trapping probabilities on the shelves of southeast Baffin Island similar to those noted above for upstream areas.

This transport process, in effect, delivers a "pulse" of bergs to the northern Labrador Shelf region. The pulse begins in February and ends (or at least decreases greatly in intensity) in July with the clearing of the coastal fast ice off southeast Baffin Island. The bergs in this pulse continue to follow the continental slope past Labrador and the Grand Banks, with diversions by currents and winds into shallower grounding areas becoming possible with seasonal clearing of the sea ice. The arrival of these bergs off southern Labrador and on the Grand Banks over the April-June period coincides with the usual "iceberg season" during which iceberg severities are at their highest levels. Most of the bergs observed each year south of 48°N (Figure 10) enter the region of interest during this period.

Given the considerable uncertainties in berg mass estimation technique, the berg mass probability distribution reported for the Labrador Shelf (Petro-Canada, 1982) is similar to the observed western Baffin Bay distributions (Marko, 1981) (Figure 11) although the absolute numbers of bergs in the former area are greatly reduced from those observed upstream.

On the basis of the satellite-tracking results, it would appear that most of the bergs which become grounded in the extensive shallow regions do not reach our areas of interest. Instead, these bergs become dissipated after multiple reflotations and re-groundings during the successive brief annual intervals of fast ice clearance.

Iceberg Deterioration

Quantitative knowledge of mass loss rates in populations of grounded and/or mobile bergs is extremely limited. Nevertheless, the significance of the mass loss process can be appreciated by noting that, even in years of exceptionally high (1000) berg counts south of 48°N, the total observed berg mass (excluding berg fragments such as growlers and bergy bits) is less than .5% of the estimated annual production of the West Greenland glaciers.

Although various schemes (Huppert, 1980; Budd et al., 1980) have been used to derive empirical estimates of berg mass loss rates, the utilized measurement techniques have been sufficiently indirect as to severely limit quantitative use. The dominant mass loss mechanism (Hanson, 1988) appears to be the wave erosion process treated theoretically (along with other mechanisms) by White et al. (1980). According to this mechanism, ocean gravity waves produce melting in a horizontal strip centred on the berg waterline by an effective forced convective process. In addition to the direct mass losses due to melting, the process also produces overhanging ice ledges which eventually fall off into the sea further reducing individual berg masses. Although expressions were also derived for the rates of ice ledge fall-off (White et al., 1980), in the absence of confirming observational data, it is difficult to incorporate such results into a single theoretical expression for mass loss rates due to all causes.

Instead, we have compared the relatively straight forward wave erosion mass loss expressions derived assuming perfectly smooth iceberg surfaces with results from two other studies: (1) empirical mass loss rate estimates made by Venkatesh (1986) for the post-reflotation period of a previously grounded berg and, (2) similar estimates made from initial and final dimensions taken from aerial planview photographs of an offshore iceberg (IIP, 1965). In both instances, the theoretical smooth wall loss rates were equal to about 5% of observed rates of total mass loss. The origins of this apparent twenty-fold discrepancy between the theoretical wave erosion induced- and observed-total mass loss rates appear to be explicable when realistic surface roughnesses and allowances for surface imperfections, ledge-falloffs and calving are included in the calculations (White et al., 1980).

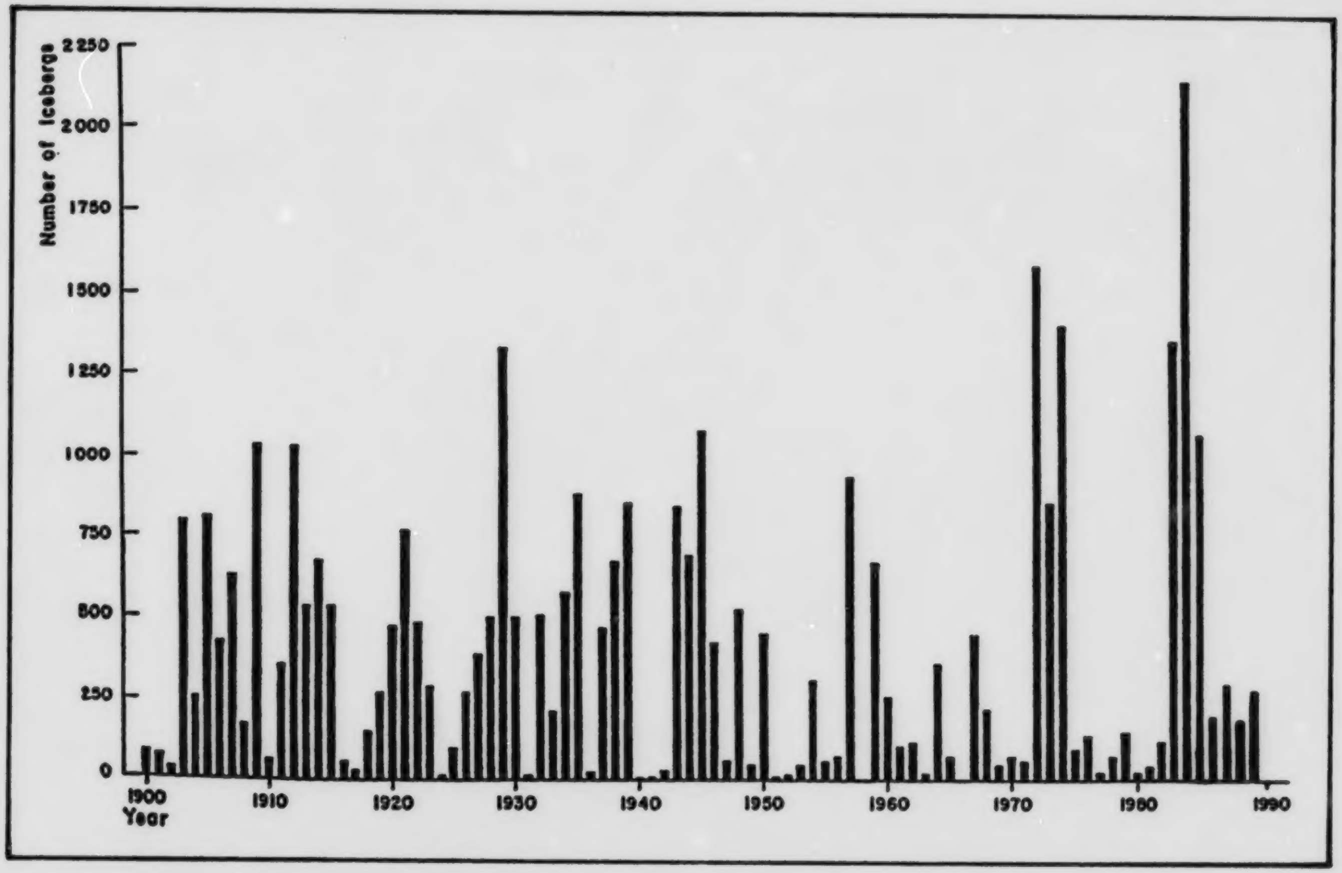


Figure 10 Annual iceberg counts as recorded south of 48°N by the International Iceberg Patrol.

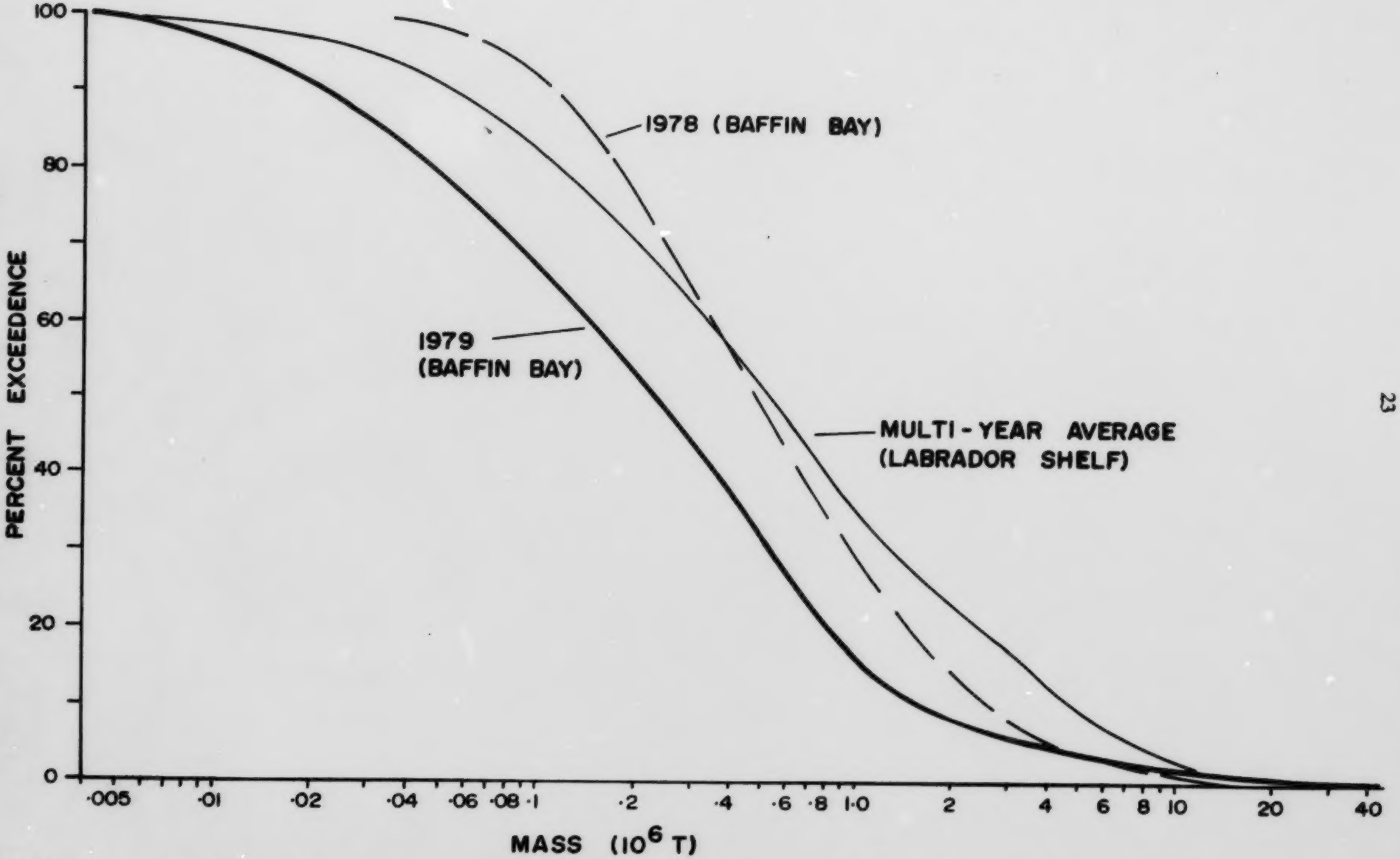


Figure 11 Exceedance percentages for iceberg masses as estimated in western Baffin Bay in 1978 and 1979 (Marko, 1981) and off Labrador (multi-year record) (Petro-Canada, 1982).

The resulting picture of iceberg production, transport and dissipation enroute to the southern areas of interest, thus, has been at least semi-quantitatively conceptualized. The key elements relevant to our considerations of possible climate induced change, include glacier dynamics, regional currents, bathymetry, water properties, and, perhaps above all, the region's extensive fields of landfast, fast and mobile ice.

3. INTERANNUAL ICE VARIABILITY

3.1 Introduction

The foregoing summaries of sea ice and icebergs have focused on the processes which govern observable environmental changes. This treatment has ignored the most striking aspect of these environmental components, namely their extremely large interannual variability. This large variability is evident in the time series of reported iceberg counts in areas south of 48°N (Figure 10). The latter time series is central to most analyses of iceberg severity variability, in spite of its considerable intrinsic uncertainties discussed above (Section 2.4). A similar, but less extensively documented variability, is also evident in the corresponding data on annual ice areal coverages and volumes in the areas of interest.

Meaningful anticipation of the effects of climate change on future sea ice and iceberg severities requires, at least as a first step, an understanding of present sources of variability. Unfortunately, to date, only relatively modest progress has been made toward obtaining such understandings. Moreover, perhaps because of the greater perceived threat to navigation, most efforts in this regard have been associated with the development of relatively short term (months) predictive capabilities relative to interannual iceberg severities. With perhaps two exceptions, most treatments (Meinardus, 1904; Mecking, 1906, 1907; Smith, 1931; Schell, 1940, 1950, 1952; Post, 1956; Corkum, 1971; Davidson, 1987; and Walsh et al., 1986) have been directed at the development of empirical relationships between various mean atmospheric- (and in some cases oceanic-) parameters and severity values usually derived from the time series data of Figure 10.

In all instances, priority was given to maximizing the correlation between observed severities and the corresponding individual or combined atmospheric and oceanic parameters. Yet, in spite of the emphasis given to establishing empirical links to such parameters, the highest correlations, or predictive successes, were reported for simple predictors based upon measured areal ice extents. Two such predictors which have been evaluated in this respect are the sea ice extents measured off Newfoundland in the spring (Smith, 1931) and in Davis Strait in January (Marko et al., 1986).

The close connection between sea ice extent and iceberg severity implied by the latter approaches to prediction has not, to our knowledge, received further investigation. Such a connection is, however, fully consistent with the process summaries of Section 2.4. Fast ice extent and duration were seen to effectively modulate the flux of bergs both from the calving glacier fronts and along the perimeter of the generalized berg circulation concentrated near the edge of the continental shelfs off Eastern Canada. Likewise, during the critical winter period, the mobile ice pack strongly determines the movement of icebergs. As well, many of the environmental factors, such as temperature and wind and current forcing, which determine sea ice extent, might be expected to affect berg numbers in a similar sense.

Finally, sea ice acts to decrease iceberg mass loss from the dominant wave erosion deterioration mechanism by damping ocean waves and holding surface water temperatures below 0°C.

These results are indicative of a fundamental difference between the sea ice and iceberg environmental components. Iceberg severity variability is strongly related to sea ice variability. The converse of this statement does not hold in most instances. Sea ice variations, except in the immediate vicinity of extremely high local berg concentrations, would be expected to be understood independently of changes in the iceberg environmental component.

As a result, we continue considerations in terms of previous and new results relevant to the origins of variability in the regional sea ice cover (Section 3.2). The results of these considerations will then be applied to iceberg severities, building on the summary in Section 2.4, to derive comparably detailed understandings of present iceberg severity variability (Section 3.3).

3.2 Sea Ice Variability

Interannual Variability - Spatial and Temporal Scales

Previous studies of the interannual variability of sea ice have been conducted over a wide range of temporal and spatial scales. For the northern hemisphere (Walsh and Johnson, 1979; Mysak and Manak, 1989; Parkinson and Cavalieri, 1989), total sea ice extent exhibits typical interannual variations of 600,000 km^2, or about 10 percent of the mean seasonal variation of 7,000,000 km^2 (Figure 12a). However, the relative degree of interannual variability in subregions is considerably larger, particularly in Baffin Bay/Davis Strait and other areas (Bering Sea, E. Greenland Sea and Barents Sea) where the wintertime ice advance is not completely limited by continental boundaries. For the Baffin Bay/Davis Strait region, the interannual variability in sea ice extent in late winter and early spring amounts to about 500,000 km^2 or about 40 percent of the mean seasonal variation of 1,200,000 km^2. The patterns of interannual variability of Arctic sea ice are dominated by an asymmetric spatial mode in which the North Atlantic anomaly is opposite in sign to the anomaly over the remainder of the polar ice cap. The time scales of the interannual variability of the Arctic sea ice appear to range from a few to many years, but are difficult to quantify given the limited (~40 year) record of observations.

Interannual variation of sea ice extent off the Canadian East Coast appears to be dominated by large amplitude fluctuations occurring on decadal, sub-decadal and inter-decadal time scales (Marko et al., 1986; Mysak and Manak, 1989; Mysak et al., 1990; Hill and Jones, 1990; Peterson and Prinsenberg, 1990; Ikeda, 1990). For the entire study area, from Baffin Bay to the Labrador/Newfoundland shelf, the most reliable sea ice extent measurements were obtained with the advent of satellite remote sensing in the late 1960's. However, less reliable estimates extend back to the early 1950's. In this larger region, the interannual variability in sea ice extent exhibits an approximately decadal cycle (see Figure 12b). Measures of sea ice extent off Newfoundland have been available for a much longer duration relative to the more northerly portions of our study area. A quantitative estimate of maximal sea ice extents (in March) for the region south of latitude 55°N has been compiled for the period 1920 to 1987 by Hill and Jones, 1990 (see Figure 12c). (Ice data are available from published sources for even earlier periods extending back at least as far as 1860; but are of poorer quality [Hill and Jones, 1990], and will not be considered further in this report).

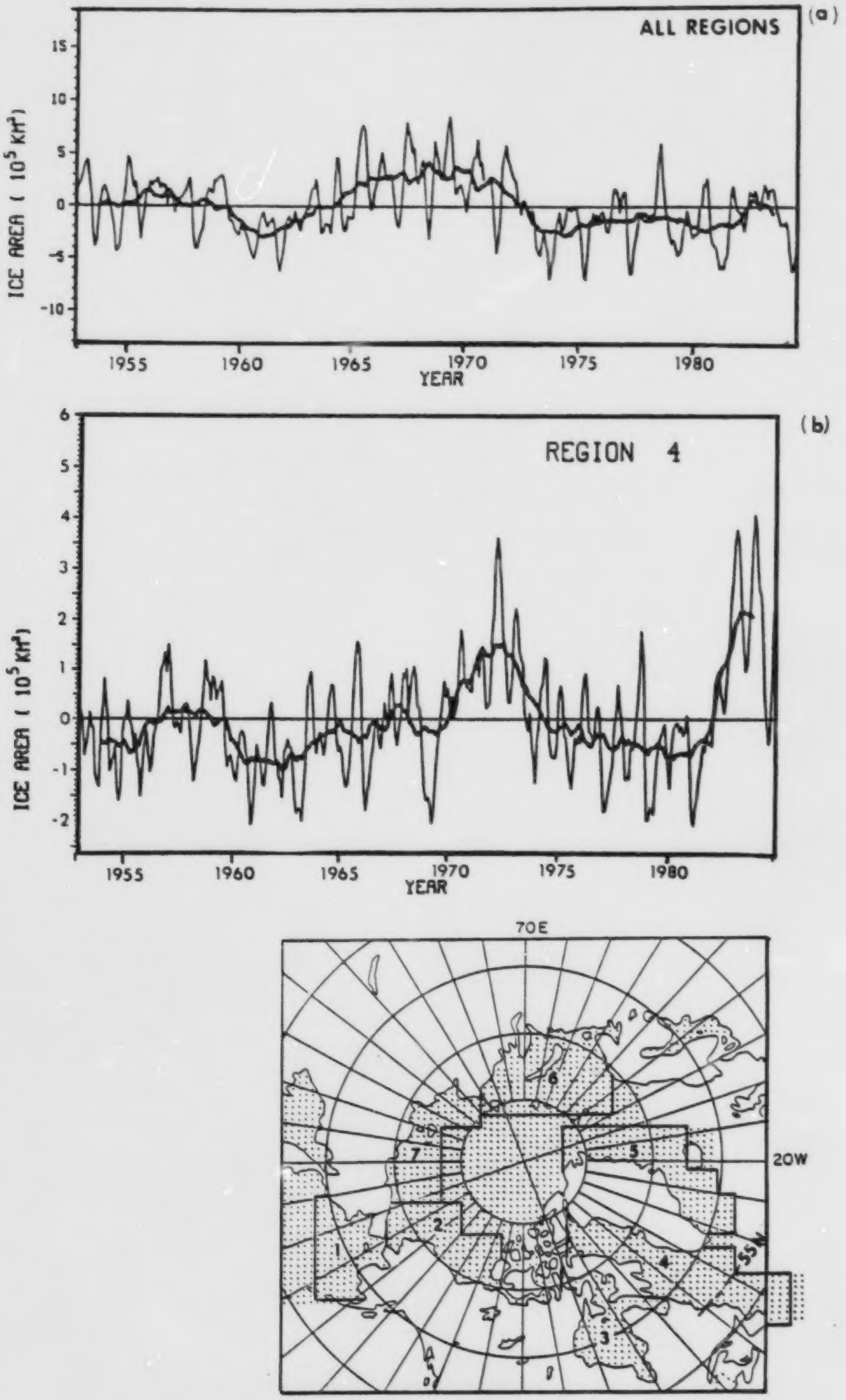


Figure 12 Interannual variability in sea ice extent for (a) the entire northern hemisphere, 1953-1985 (Mysak and Manak, 1989); (b) Baffin Bay/Labrador Sea, 1953-1985; (Mysak and Manak, 1989); continued on next page...

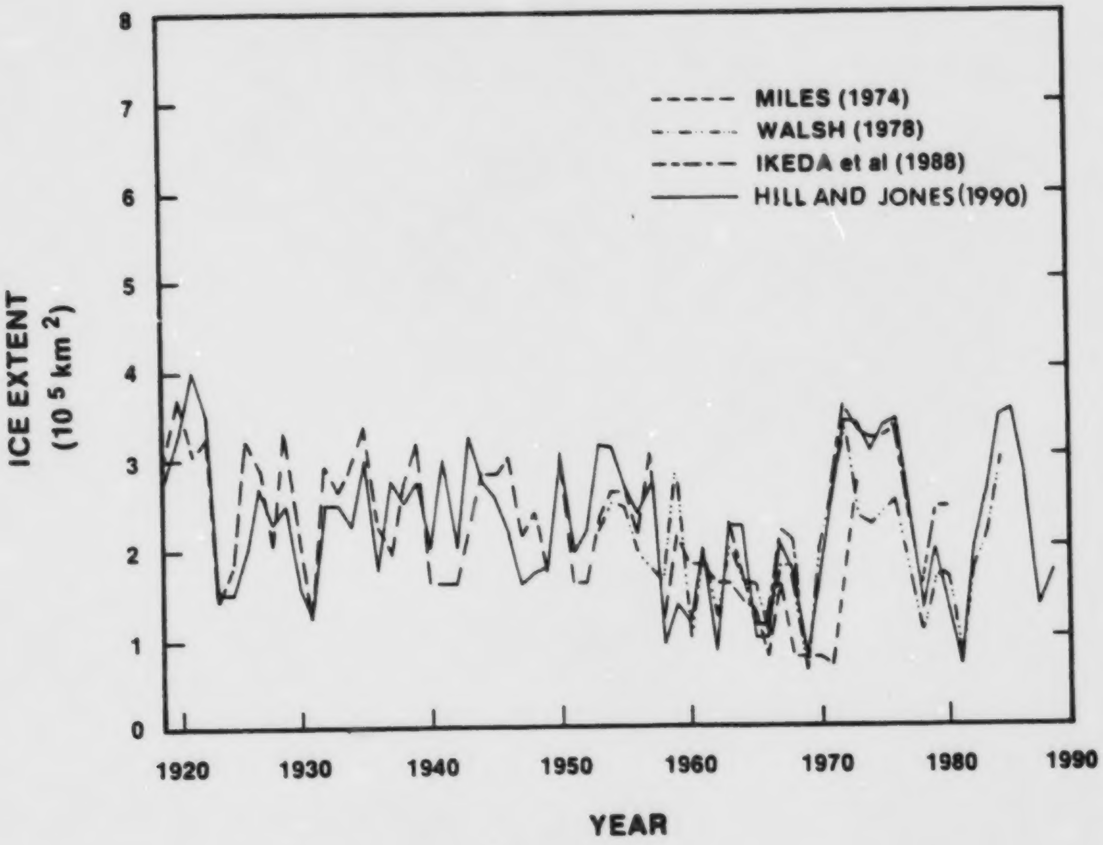


Figure 12 Interannual variability in sea ice extent for (c) South Labrador/Newfoundland (south of 55°N; 1920-1988; Hill and Jones, 1990).

From the 1920-1987 data set, there is evidence of approximately decadal scale oscillations in sea ice extent combined with longer period variability. The longer term, sub-decadal, variations are evident in the generally low sea ice extent of the 1960's and generally high extent of the 1930's to early 1940's. Estimating the periodicity of the long-term variations is a difficult task, given the comparatively short (68 year) record available.

The causes of the large amplitude variability in sea ice extent are not immediately evident on either decadal or longer time scales. Many investigators have reported statistically significant correlations of sea ice extent with meteorological and oceanographic parameters, and in particular, winds and air temperatures (Symonds, 1986; Marko et al., 1986; Ikeda et al., 1988; Peterson and Prinsenberg, 1990). It is recognized, however, that at least some of the observed similarities in variability may represent a common response of coupled variables to external forcing.

Linkages to External Parameters

Sea ice extent has also been shown to be correlated, in a statistically significant manner, to even larger scale external parameters, including the North Atlantic Oscillation (NAO) Index (Mysak and Manak, 1989). The NAO represents a complex internal mode of natural variability with decadal-like time scales, which may be associated with other internal atmospheric-ocean phenomena such as the El Nino/Southern Oscillation.

One hypothesis for the decadal-scale variability in sea ice extent is related to variations in solar radiation (Hill and Jones, 1990). This hypothesis arises from approximate coincidence of the observed decadal variations in sea ice extent with the sunspot cycle. Hill and Jones (1990) report a statistically significant correlation coefficient of 0.45 between Newfoundland ice extent (1920 to 1988) and a derived index of sunspot activity. The correlation improves to a value of 0.52 when comparisons are limited to the west phase of quasi-biennial oscillation (QBO). While the proposed causal relationship appears to be a straightforward case of climatic forcing, the change in solar irradiance is too small to directly account for the variations in temperatures and ice extent. The climate forcing attributed to sunspots is small (0.24 Wm^{-2}) in comparison to other external climatic forcings such as changes in atmospheric aerosols due to volcanoes (~1.0-1.5 Wm^{-2}) and greenhouse forcing (2-2.5 Wm^{-2}) (Hansen and Lacis, 1990). While Hill and Jones (1990) did not offer detailed mechanisms to account for the significant observed correlations, they suggested that intermediary linkages in the atmosphere-ice-ocean system, such as large-scale wind patterns, may play a role in enhancing the comparatively weak climatic forcing associated with solar irradiance variability.

Recently, other mechanisms have been advanced to account for sea ice variability based upon complex atmosphere-ice-ocean feedback mechanisms operating on very large scales. One hypothesis of this nature (Ikeda, 1990) linked local weather conditions off the Canadian East Coast (and the resulting variations in sea ice extent) to unusual atmospheric pressure patterns over the Arctic Eurasian Basin. Ikeda (1990) presents a mechanistic two-level ocean model driven by buoyancy flux and wind stresses, which suggests that the Arctic-Atlantic system resonates with the variable wind stresses at a decadal time scale. A conceptual air-ice-ocean model that includes the oceanic model response as well as positive feedback between air and ice-ocean systems shows that weak decadal variations in external forcing can be amplified in the air-ice-ocean system. Mysak et al. (1990) followed the same conceptual framework (developed by Kellogg, 1976) to develop a negative feedback system linking sea ice extent in the Greenland Sea (and subsequently in the Labrador Sea through ocean advection) to oceanographic properties in the Arctic Ocean and Greenland

Sea which are related, in turn, to Arctic cyclogenesis and precipitation over northwestern Canada and the resulting runoff into the Western Arctic Ocean. The combination of this complex set of feedback linkages would result in a pseudo-cyclic interannual variation with very long period, perhaps on the order of 30 years. Because of its complexity, and the very large spatial area involved, the validity of the proposed feedback system over the Arctic-North Atlantic Ocean basins is very difficult to examine quantitatively over several cycles due to data limitations.

Relationship to Internal Parameters

Several early studies (Dunbar, 1972; Barry and Keen, 1978 and Crane, 1978) have examined trends and details associated with the variability in the advance and retreat of both fast and mobile ice in the northern portion of our study region (i.e. Davis Strait and Baffin Bay). As indicated in Section 2.1, some of this work (Barry and Keen, 1978) has attempted to establish causative links between the characteristics of ice change (i.e. rates of retreat/advance) and details of atmospheric features, such as the position of the 700 mb pressure trough. Nevertheless, the most quantitative and directly applicable studies of variability in the regional ice cover have been carried out by Symonds (1986) and by Ikeda et al. (1988) as part of an ongoing program to model the ice off southern Labrador and on the Grand Banks.

The latter work was initiated to address the apparent failure (Hibler and Walsh, 1982) of a two-level Hibler ice model (Hibler, 1979, 1980) to correctly simulate observed mean ice extents and interannual variability in the North Atlantic region, including the Labrador-Sea and -Shelf regions. Ikeda et al. (1988) surmised that this failure of the model, which had previously reproduced observed extents and variabilities in the Arctic Basin, was attributable to an underestimation of the oceanic heat fluxes due to the neglect of convective over-turning in the water column.

Tests of the effects of underestimating heat fluxes were carried out (Ikeda et al., 1988) using an eleven-year record of sea ice areal extent data for the region south of 55°N. These data were used to "calibrate" and test a two-level Hibler Model simulation (Hibler, 1979) which extended north to 60°N in order to account for the advection of ice into the study region. The "calibration" involved selecting an interannually invariant, non-zero oceanic heat flux by optimizing agreement between simulated and observed ice extents for the 1971 ice year. This heat flux, 35 W/m^2, was significantly larger than the 2 W/m^2 value assumed by Hibler and Walsh (1982), thus accounting for the previously, larger than observed, simulated extents. The larger flux was assumed to be present only during the first 3 months of the December-June modelling period with subsequent fluxes set equal to zero. Winds were interpolated on a 25 km x 50 km grid from six hourly geostrophic winds computed by the Canadian Climate Centre. Air temperature and dew point fields were also interpolated from daily averages at six coastal stations and Ocean Weather Station Bravo (Figure 1), including an attempt to account for the bias toward coastal station data. Cloud data were taken from the Station Bravo observations and assumed to be applicable through the modelling regions. Climatic values of the regional currents (Lazier, 1982) and of the initial (December) ice free water temperatures were assumed. Atmospheric heat exchanges were determined using the bulk aerodynamic formulae of Smith and Dobson (1984).

The results shown in Figure 13 showed good, although marginally statistically significant, agreement with observations for the first four months of the simulation period (December-March). Sensitivity and other tests with model parameters suggested that, during this

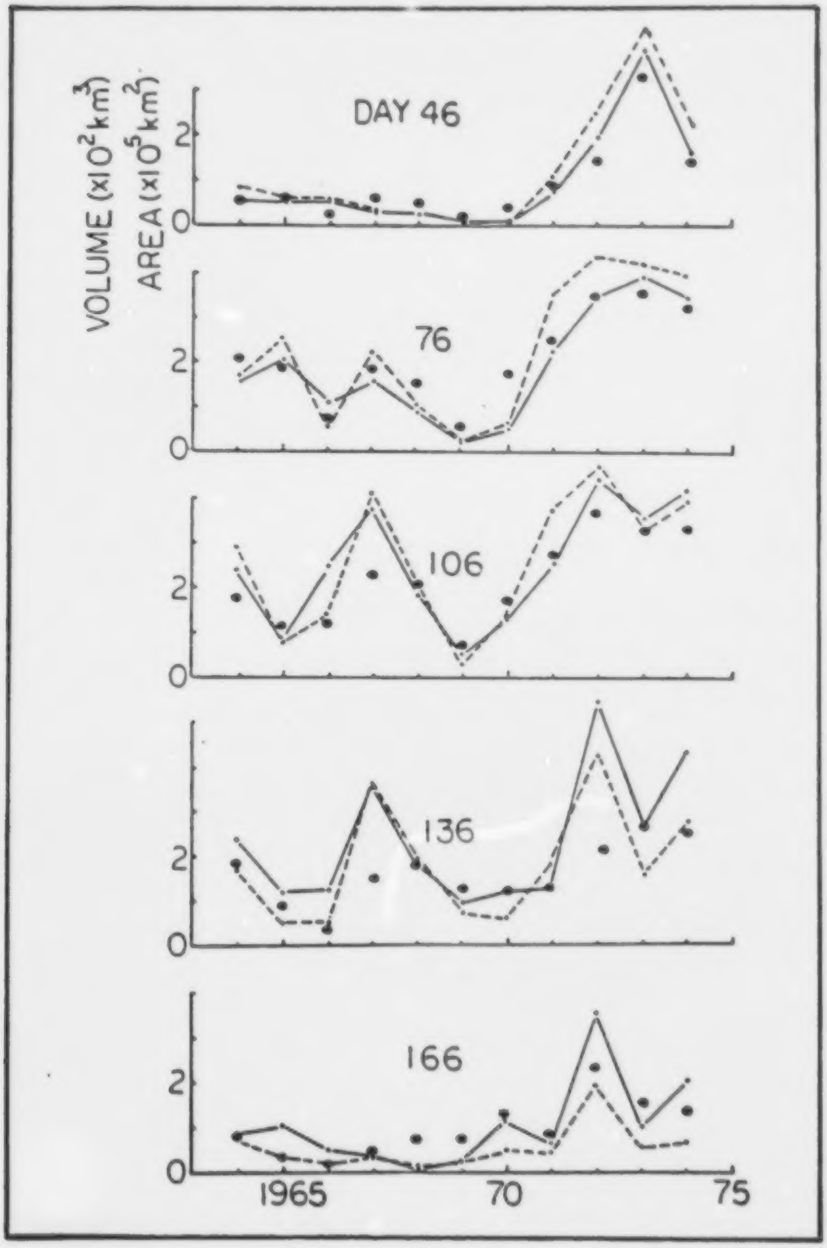


Figure 13 Ice covered areas (dashed lines) and ice volumes (solid lines) south of 55°N on approximately (top to bottom) Jan. 15, Feb. 14, March 15, April 15 and May 15, as derived with the Ikeda et al. (1988) model. Also shown are the actual observed ice covered areas as solid dots.

period, variability in areal ice extent was largely determined by air temperatures and wind speeds. Ice extent increased with increasing wind speeds and falling air temperatures.

Increased advection of the ice by larger alongshore wind components was of lesser significance to interannual variability. In general, the implication of the December-March results was that the role of the ocean was of secondary importance in determining ice extents relative to the much larger energy exchanges with the atmosphere.

Nevertheless, as inspection of Figure 13 indicates, the earlier success of the Ikeda et al. (1988) simulations began to break down for periods later than March 15. More specifically, in the months of April and May the mean errors in the annual simulated ice extents became comparable to the standard deviation of the observed annual spring ice extents. Several possibilities were cited for this breakdown (Ikeda et al., 1988). The most likely of these possibilities, in our opinion, is the model's neglect of advection of ice from more northern areas into the modelling region south of 60°N.

To examine this possibility further, interannual variations in the Labrador spring ice extent were compared directly with corresponding earlier variations of ice extent in appropriate upstream areas. The time and location of the upstream ice comparisons were chosen to be, respectively, the end of January and the non-coastal portion of Davis Strait between 61°N and 67°N. Assuming a 15 km/day average southward drift velocity, the ice found in Davis Strait at that time moves across the 55°N northern boundary of the Labrador ice extent measurement area between, roughly, March 15 and early May. This timing would suggest that the identified late January Davis Strait ice could be a dominant presence in the south Labrador region during the anomalous April, May and June periods. This circumstance provided at least part of the motivation for the earlier (Marko et al., 1980) definition of the Davis Strait Ice index for comparison with spring and summer iceberg severities south of 48°N. This index represented the normalized areal extent of ice on concentrations equal to or greater than three-tenths present in the specified portions of Davis Strait at the end of January.

Unfortunately, lower resolution and inconsistencies in some of the pre-satellite era data obtained for the years prior to 1972 precluded fully reliable comparison of upstream data with all of the 1964-1974 Labrador ice extent data considered in the Ikeda et al. (1988) work. Instead, rigorous comparisons were confined to the 1972 to 1989 period when the availability of satellite imagery allowed reasonably accurate extraction of end of January ice extents. In each case, areal coverage was computed either directly from NOAA AVHRR images or from the annual end of January ice charts prepared by the U.S. Naval Fleet Weather Centre and Environment Canada. The Labrador data used in the comparisons were taken from the sum of extents reported for April 15 of each year by Peterson and Prinsenberg (1990) for the period 1972-1988 (and in 1989 by direct measurement from the U.S. Naval Fleet Weather Centre ice charts) in each of three zonal strips extending off Labrador and Newfoundland between, respectively, 45° and 47°N, 49 and 51°N, and 53° and 55°N.

A plot of the annual south of 55°N ice extent sums relative to corresponding end of January and Davis Strait ice extent indices (areal extents of ice in current ratios greater than two-tenths, normalized to 150,000 km^2) indicates a near-linearity (r=.9) in the relationship between the quantities in all but two years (Figure 14). The latter notable exceptions corresponded to data obtained during 1975 and 1981 when very strong anomalous easterly wind components were reported in the Labrador Sea in April and March, respectively, on the basis of reported monthly mean pressure fields (IIP, 1976, 1982). Similar March and/or

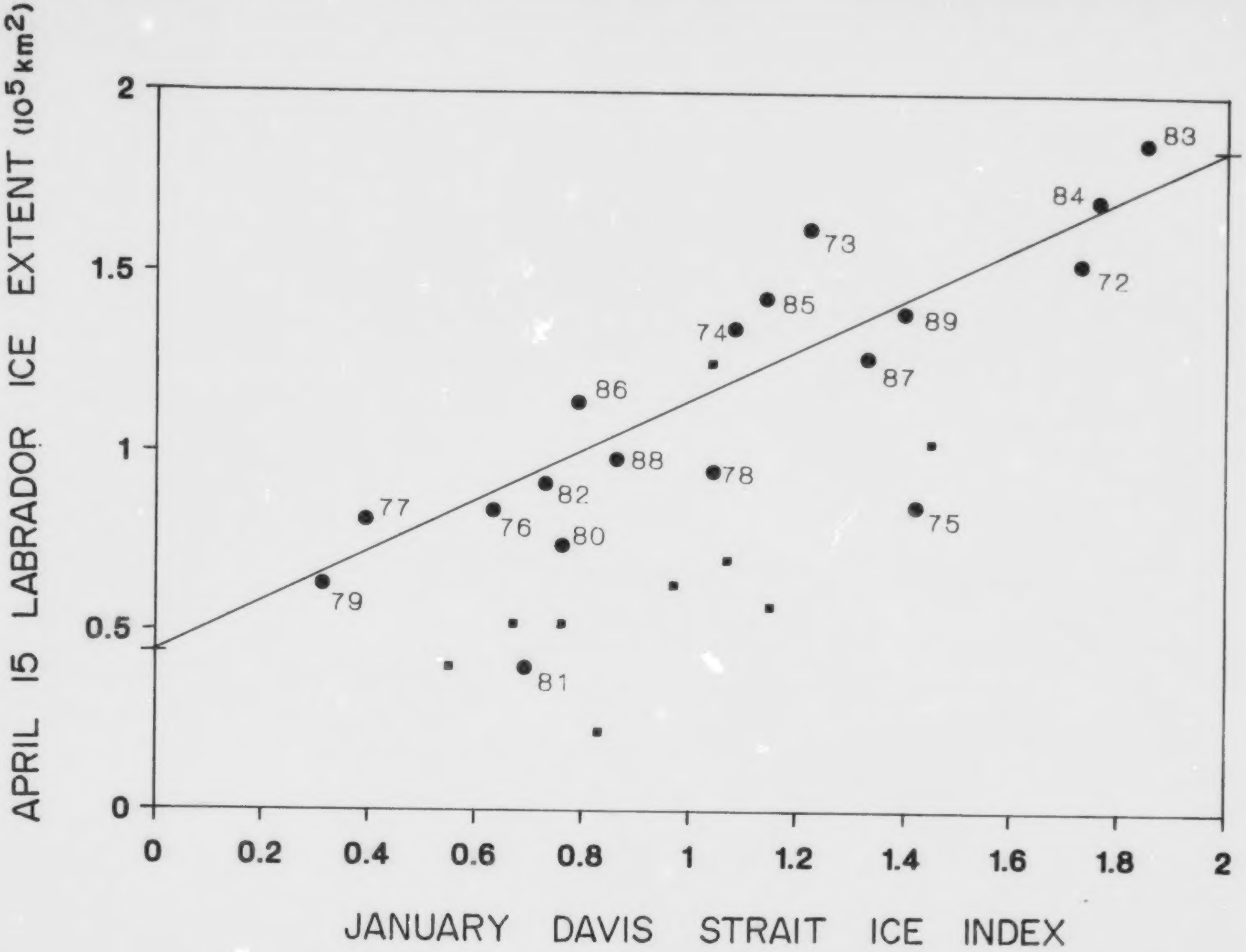


Figure 14 Observed relationships between corresponding annual April 15 Labrador ice extents in three 2° latitude strips between 47° and 55°N (Peterson and Prinsenberg, 1990) and the January Davis Strait Ice extent index (Marko et al., 1986). The labels identify the corresponding solid circle data points for the years 1972-1989. The included line represents a linear regression (r=.9) derived from the 1972-1989 data set, excluding the 1975 and 1981 data points. Also plotted (unlabelled small square points) are the values from 1963-1971, obtained from Walsh (1978) for the period 1963-1971 (see text).

April wind configurations were also noted during the years 1972, 1976, 1987 and 1989, also associated with data points which lay below the indicated regression line.

The additional unlabelled data points denoted by small squares in Figure 14 are representative of years in the pre-satellite 1963-1971 era associated with greater uncertainties in Davis Strait ice extent estimates. South of 55°N ice extent values for these years were obtained, again, from Peterson and Prinsenberg (1990) while the Walsh (1978) compilation was used to obtain the Davis Strait data. Uncertainties in the latter data are associated with: the underlying relative spatial and temporal detail available about ice edge position; the use, at various times, of data from two or three different ice monitoring agencies using slightly different classification categories; and the coarser spatial resolution of the compilation data (ice concentration given on a one degree latitude grid). When translated into an appropriate ice boundary position and integrated over the measurement region, the data points, with the exception of single high and single low outliers, show a similar interdependence between the upstream and downstream ice extents. The low-lying point did correspond to an interval of greater than average easterly wind conditions during March 1966. This interdependence is reasonably well represented by a linear relationship with roughly the same slope as that associated with the post-1971 data, but with a lower intercept on the Labrador ice extent axis. This difference could be associated, in part, with either systematic errors in the earlier area estimates but, as well, could be indicative of a greater tendency toward intervals of strong easterly winds and/or thinner ice during the earlier period. In any case, the consistency of linear relationships between comparable sets of winter upstream and spring downstream ice extent data suggests the existence of an underlying variability mechanism(s) which is relatively independent of the significant changes in the regional climate which can be associated with the larger 1963-1989 period (see Section 4.3).

The apparent difference between Labrador spring ice extents corresponding to years, alternatively, with and without significant easterly spring wind components is consistent with numerous past and more recent observations (Carsey et. al, 1989) of wind-induced surges and retreats in ice boundaries. In particular, easterly wind components tend to compress the pack ice, forcing a westward retreat of the ice boundary. Nevertheless, the role of easterly winds in rapidly reducing the ice extent is not a simple process, depending on the complex mechanical and dynamical properties of mobile pack ice. For example, early occurrence of easterly winds, when the ice tends to be thicker, may not be as effective in producing changes as later season winds. The results of Figure 14 can thus be interpreted in terms of two factors acting as determinants of springtime ice extent off Southern Labrador. One of these, the late January sea ice extent in Davis Strait, appears to set a maximum limit for ice area to the south of 55°N in mid-April. The other factor, easterly (or onshore directed) winds, subsequently determines the degree of reduction from the maximum value.

The one-sided role of the Labrador spring winds in this process appears to be attributable to the inherent opposite signs of the feedbacks associated with the alternative easterly and westerly wind components. Easterly winds both compress and reduce ice extent and increase ice meltback by moving warmer offshore surface water toward the ice and by increasing overall solar energy absorption by reducing fractional surface coverage by high albedo ice. This positive feedback enhances the ice extent reduction process and contrasts with the westerly wind component situation in which the initial eastward shifts of the ice boundary are generally limited (negative feedback) by the increased meltback produced by closer proximity to warmer offshore waters. The ice advected and lost to the offshore is replaced by new ice growth throughout the winter and early spring off Labrador.

Taken together, the results obtained by Ikeda et al. (1988), and in this study, suggest that present day interannual variability in the Labrador Sea-Grand Banks winter ice extent is primarily determined by ice growth off Labrador and subsequent southward advection of ice. In spring, the advection of ice originating in Davis Strait into the S. Labrador-Grand Banks area, and the effects of Labrador March/April winds appear to become important. Note, however, that other more subtle processes also may be operative in the spring period, including neglected oceanographic effects. One such possibility is stratification in the upper ocean layer due to partial ice melt or freshwater discharged into the area from upstream sources such as Hudson Bay. The effects of the ocean stratification, which could enhance sea ice production within the Labrador Shelf region, are a matter of ongoing research (Yao and Ikeda, 1990; Ikeda, pers. comm.). Nevertheless, on the basis of the existing data sets, the January Davis Strait ice extent appears to provide a reliable indicator for the upper limit of the southern regional ice extents during the following spring period.

Understanding this apparent connection between upstream, midwinter- and downstream spring-ice extents would appear to be a necessary step in any attempt to assess the likely effects of climate change on spring ice conditions south of 55°N. On the basis of the modelling results obtained off Labrador (Ikeda et al., 1988) and the correlations between Labrador and Davis Strait ice extents, one could expect to link variations in the January Davis Strait ice extent to atmospheric factors using a simple, atmospherically driven ice model. The value of the oceanic heat flux would be selected once advection of ice from upstream areas is properly included. In this instance, however, the modelling area would have to include part or all of Baffin Bay and the absence of thick ice in the region prior to freeze-up cannot be assumed.

Initial tests of January wind and ice extent correspondences in Davis Strait over the 1972-1987 period have shown northerly wind forcing to be a likely major source of ice extent variability. In particular, high Davis Strait ice indices were recorded in each of the eight years during the period in which the mean monthly northerly wind component (at 60°N, 55°W) exceeded 13 knots. On the other hand, the remaining high ice extent year, 1985, was associated with strong (17 knot) easterly mean winds. Such deviations from a simple wind-driven determination of Davis Strait ice extents are not unexpected. One source of complexity and deviation could be associated with ice exchanges with Hudson Strait. Other possible complications associated with similar exchanges with more northerly side channels will be discussed below.

Sea Ice Contributions to Northern Baffin Bay from Adjoining Channels

Contributions to variability in the apparently important (to spring sea ice conditions south of 55°N) January Davis Strait ice extent parameter may be associated with the significant exports of ice into northern Baffin Bay by Lancaster, Smith and Jones Sounds. Such externally generated ice not only enters Baffin Bay during the summer and early fall as remnants of the disintegrating Arctic Archipelago ice cover, but actually can increase in volume during the late fall and early winter months when such exports could affect the January Davis Strait ice extents.

In the most thoroughly documented channel, Lancaster Sound, variability in exported ice volume is in large part associated with the stabilization of alternative fast (immobile) ice edges at one or another of a small number of widely separated locations along the length of the channel. On the basis of the observed characteristics (Appendix A) of these edges and of the mobile ice drift, the volume of the ice exported into Baffin Bay should be very roughly proportional to the distance between the prevailing ice edge and the channel junction with Baffin Bay. The calculations, discussed in Appendix A, suggest that over the November-January period, when channel exports might be expected to have influence on the end-of-January ice edge position downstream in Davis Strait, the two extreme eastern and western positions of the ice edge could correspond to a 75 km^3 net difference in total exported ice volume. This volume is almost 20% larger than the standard deviation of the estimated annual volumes of ice found in Davis Strait in late January. Similar shifts have been noted in Smith and Jones Sounds (Marko, 1981), but data are presently insufficient to quantify the magnitudes of the corresponding ice export variability or to establish phase relationships, if any, with Lancaster Sound processes.

In the absence of detailed information on the annual variability of the November-January ice edge position, evidence for the possible downstream significance of channel exports is presently circumstantial and confined to the observation that the lowest yet recorded (in 1979) value of the Davis Strait ice index (.31) followed the earliest observed (early November 1978) stabilization of an annually permanent ice edge at an extreme eastern position, insuring near-minimal seasonal ice exports. Should such a linkage be confirmed by detailed studies, the task of establishing linkages between downstream (south of 55°N) ice extents and large scale atmospheric influences could be complicated by possible dependencies of side channel exports on currents, ice strength and thickness and other factors. Such dependencies could be very sensitive to the air-temperature and -pressure changes which might be expected to accompany global warming.

3.3 Iceberg Variability

Relationship to Variability in Sea Ice Extent

The physical linkages and statistical correlations discussed in Section 3.1 suggest that a useful working relationship may exist between annual variations in ice conditions and corresponding changes in iceberg severities in our areas of interest. Such a relationship would provide a natural starting point for developing an understanding of the absolute and relative importances of the three primary generic sources of present day iceberg severity variability, namely changes in berg-production (calving), -transport and -deterioration (mass loss).

Unfortunately, progress in this regard has been limited by the inadequacies of the iceberg databases and, in particular, by large uncertainties in the most consistently recorded iceberg parameters, namely the annual numbers of icebergs observed south of 48°. These uncertainties (as discussed in Section 2.4) are difficult to quantify without detailed reviews of the technologies, procedures and levels of effort used in each annual survey. Nevertheless, it is not unreasonable to assume that at least in the era preceding the institution of regular all-weather radar survey flights (1983), errors in the individual annual count values could be as large as 100% or more. The possibility of such large errors in at least some of the annual data points greatly reduces the applicability of simple regression techniques. Nevertheless, the use of such techniques in conjunction with the berg count database has consistently (Smith, 1931; Walsh et al., 1986; Davidson, 1987) shown a tendency for higher berg counts to be associated with preceding atmospheric configurations which produced above-average northerly or northwesterly wind components in the study area. However, thus far it has not been possible to use these results to establish the extent to which frequent inconsistencies between observed and expected berg counts are associated with neglected physical sources of variability as opposed to errors in the underlying iceberg count database.

Minimizing chances of possible neglect or misidentification of physical sources of variability requires explicit accounting for berg count uncertainties in establishing correspondences with other environmental parameters. Unfortunately, in most cases, past attempts (Davidson, 1987; Walsh et al., 1986) to account for such defects have been effected after the derivation of empirical regression relationships by, for example, classifying the observed and relationship-predicted count values into upper, middle and lower tercile categories.

One three-level scheme (Marko et al., 1986) did carry out coarsening classifications prior to comparisons with the predictor variable by rather arbitrarily (but in accord with general IIP hazard assessments) defining berg counts below 201, between 201 and 599, and above 599 as low, intermediate and high berg severities, respectively, for the study period.

In view of the correspondences between berg severities and sea ice extents noted both in the latter work relative to the Davis Strait ice index (Marko et al., 1986) and by Smith (1931) with regard to spring ice off Newfoundland, we have examined the empirical relationship between the annual berg counts and the previously discussed spring (April 15) ice extents south of 55°N. The resulting plot (Figure 15) of 27 years (1963-1989) of ice extent and iceberg count data suggests an obvious segmentation in which berg counts above and below, very roughly, 200 correspond to south of 55° ice extents larger and smaller, respectively, than approximately 1.2×10^5 km^2. Iceberg severity class lower boundaries of 200 (intermediate) and 600 (high) as specified by Marko et al. (1982) are included in the Figure.

Excluding the single outlier noted on Figure 15, the data displayed a generally good linear relationships above (r=.65) and below (r=.74) the 1.2×10^5 km^2 ice extent threshold. In both regions, the deviation of most of the data points from the regression line were not greatly in excess of the uncertainties in the count values.

On the basis of these correspondences, it would appear that the spring ice extent off southern Labrador and on the Grand Banks is an excellent indicator of the corresponding annual iceberg severity, at least to resolution of the three defined severity categories. More sea ice clearly tends to presage more icebergs and there is strong evidence for a threshold ice extent for years with high iceberg counts.

Taken together with the discussions of the possible sources of the Labrador/Grand Banks ice variability in Section 3.2, these results suggest the existence of, at least, a good correlative relationship between berg severity and a combination of upstream (Davis Strait) January ice extents and local (south of 55°N) spring winds. This possibility was confirmed in a comparison of the 1972-1989 three-level berg severities against the original Marko et al. (1986) Davis Strait ice index thresholds. All of the four significant errors in ice index-based forecasts corresponded to lower berg severities than anticipated on the basis of upstream ice extent alone, coincident with the occurrence of easterly wind components capable of forcing springtime reductions in south of 55°N ice extents.

On the basis of the discussions of Section 3.2, a viable interpretation of these results is that the January ice extents in Davis Strait set an upper limit for both the south of 55°N spring ice extents and the south of 48°N iceberg numbers in the spring-summer months. In both instances, actual extents and berg numbers are further determined by the degree to which the southern spring winds produce simultaneous reductions from the respective annual upper limits.

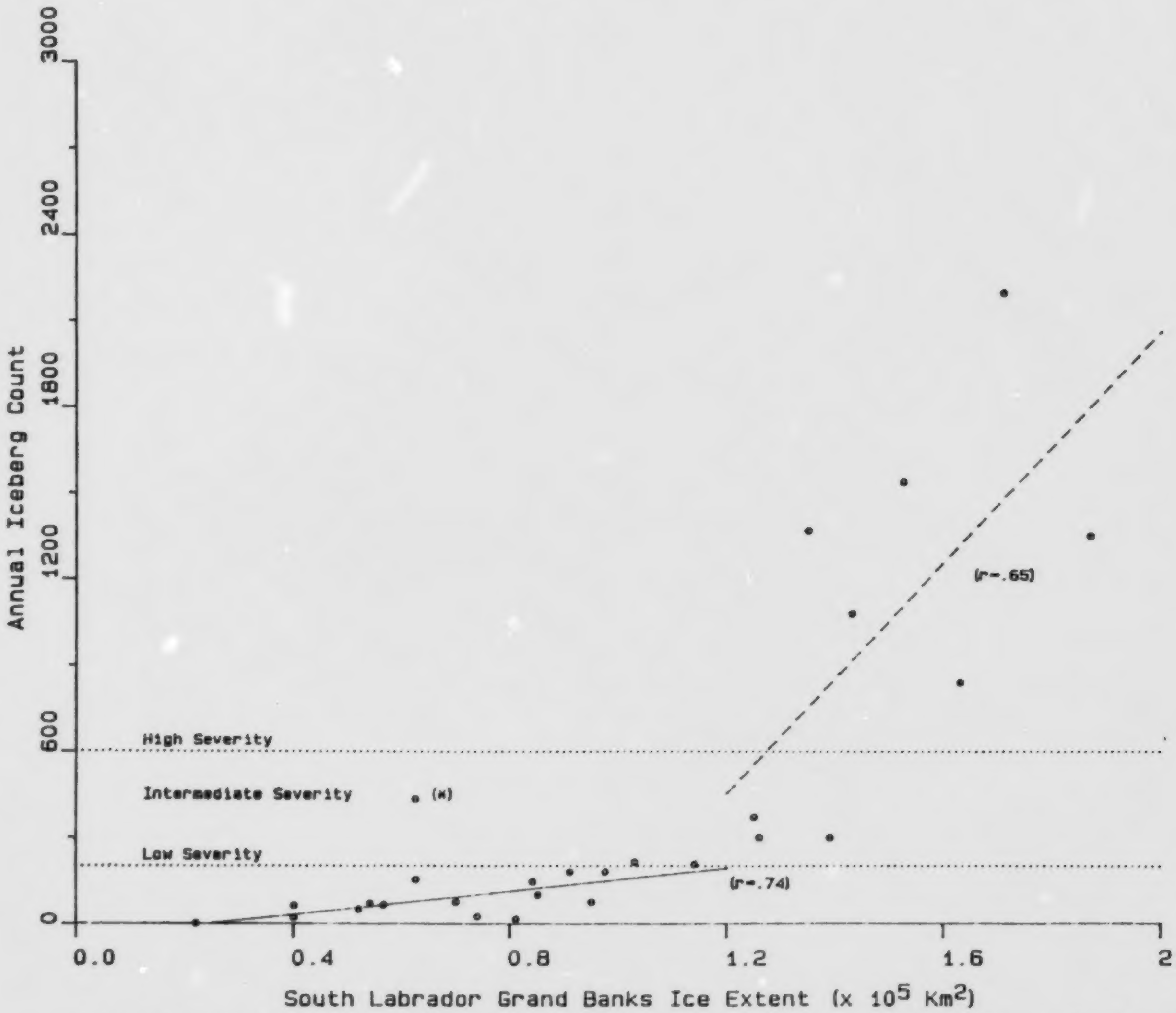


Figure 15 Iceberg counts (1963 - 1989) as a function of the April 15 ice extent off S. Labrador and Newfoundland, as estimated from three 2 degree latitude strips between 47 and 55 N. The figure also includes linear regression lines in both the low and high extent regions and the boundaries of the high, low and intermediate iceberg count categories described in the text. An outlier point corresponding to 1967 values is indicated by an asterisk (*).

The magnitudes of the possible reductions in sea ice extent can be very roughly estimated from the separation between the highest and lowest annual south of 55°N ice extents occurring at a given approximate Davis Strait ice index value in Figure 14. Similar judgements on corresponding wind-related reductions in berg numbers are compounded by both the much larger uncertainties in the count data and, undoubtedly, by other non-wind-related factors, such as year-to-year differences in incident berg fluxes from more distant upstream areas. Fluctuations of the latter type might be expected to be linked to interannual differences in calving rates and upstream grounding probabilities. The results of Figure 15, together with recognition of spring wind effects, suggest that, to some degree, such fluctuations contribute primarily to interannual differences which usually fall within the limits of the three defined berg severity categories and, perhaps, to year-to-year variations in the duration and peak timing of the iceberg season.

With regard to the latter, intra-annual aspect of berg severity variability, a review of the 1970-1989 iceberg data indicated little consistency in the observed relationships between overall annual severity and the timing of the peak in south of 48°N berg numbers. In general, larger annual severities increase the numbers of bergs arriving in all months during the January-July period. Because of this circumstance in high severity years, "early arriving" bergs are noted, but, in the period studied, peak berg fluxes were confined, with one exception to the April-June months with the annual maximum occurring most frequently (38% probability) in April.

Interpretation of the above results with regard to further clarification of the origins of the present sensitivity of the Labrador/Grand Banks region to climatic change must allow for two alternative possibilities which, respectively, attribute the cited correlations to either or both: (1) similar sea ice and iceberg responses to a common environmental change; and (2) a direct causative relationship.

In our view, however, the consistency of the observed correspondences and the evidence, in years such as 1975, of rather sudden transitions from imminently high to low iceberg severities under strong easterly wind components are most compatible with a causal variability mechanism. A specific candidate for such a mechanism could be based upon sea ice control of iceberg dissipation or mass loss through the capacities of pack ice for holding upper layer water temperatures below the 0°C meltpoint of berg ice and for damping ocean gravity waves.

First Order Model of Iceberg Deterioration as a Function of Sea Ice Extent

To test this hypothesis, rough calculations were carried out for the mass losses incurred by bergs arriving at the location 48°N, 48°W at varying times during the April-June "iceberg season", assuming that bergs moved along the 1000 m bathymetry contour at 15 km/day. These calculations were carried out for ice extents corresponding to the minimum, mean and maximum extent configurations given by the Naval Oceanography Command Sea Ice Climatic Atlas (NOC, 1986). These configurations roughly corresponded to low, intermediate and high south of 55°N ice extents as defined by Marko et al. (1986) to correspond with observations in years with both, respectively, low, intermediate and high January Davis Strait indices and negligible spring easterly wind components.

Because of our assumption that the bergs continue to follow the 1000 m depth contour, the different ice configurations, in effect, determine the duration of open water travel prior to arrival at the impact point. Losses were calculated as a function of arrival date for two idealized rectangular parallelepiped icebergs, using the principles outlined in the "Iceberg

Deterioration" portion of Section 2.4. The effects of erosion were assumed to be confined only to the open water portions of the trajectories and when surface water temperatures rose above 0°C. Mass loss rates were computed using the smooth surface wave erosion estimates from White et al. (1980) and an empirical factor of 20 multiplier derived from comparisons against observed mass loss rates (Section 2.4). Initial (while still in the sea ice) berg masses of 2 million and .5 million tonnes were assumed along with corresponding initial dimensions of 200 m x 100 m x 100 m and 126 m x 62.5 m x 62.5 m. Daily losses were calculated on a monthly basis and adjusted to scale downward all linear dimensions and, hence, loss rates accordingly. The bergs were assumed to move with their broadest horizontal dimension perpendicular to the dominant wave direction. The temperature and waveheight parameters required for the loss calculations were derived from a review of seasonal open-water oceanographic observations in the vicinity of the regional continental slope. These parameters were:

Waveheight, h_w:

$h_w = 0$ within the ice boundaries at all times,
$h_w = 1.3$ m outside the ice boundaries in the March through June period,

Sea surface temperature, Tss:

Tss ≤ 0.0°C within the sea ice boundaries at all times and in all areas prior to April,
Tss = 1.0°C outside ice boundaries in April,
Tss = 2.0°C outside ice boundaries in May,
Tss = 3.0°C outside ice boundaries in June.

The calculated results for the arriving berg masses (Figure 16) show that the smaller berg reaches the impact point with significant mass only during the month of April under both of the two assumed lower ice extent configurations. Even the larger 2 million tonne berg only survives into June under the maximum ice extent configuration. In effect, the seasonal ice extent sets a minimum size requirement for berg survival through the course of the annual April-June iceberg season. The curves in Figure 16 suggest that, for minimum ice extent, the minimum sizes required for survival through April, May and June, respectively, would be .75, 2.5 and 4.5 million tonnes. On the basis of the exceedance data (Figure 11) and assuming constant monthly berg fluxes (in the pack ice), this requirement, of itself, would produce a larger than factor of 4 difference between the overall seasonal totals of bergs reaching 48°N, 48°W under the alternative minimum and maximum ice extent configurations. This figure is comparable to the factor of three difference between the counts associated with respectively the upper and lower ends of the previously discussed low and high berg severity ranges.

It is significant to note that the roughly 1.2×10^5 km^2 threshold in the empirical relationship (Figure 15) between the berg numbers south of 48°N and the defined south of 55°N ice extents corresponds to an offshore pack ice boundary lying close to the 1000 m depth contour marking the approximate axis of the offshore branch of the Labrador Current. This result suggests that a given change in total ice extent has less impact on seasonal iceberg numbers if the pack ice edge is inshore of, as opposed to in or beyond the strong current regime. In the absence of more detailed data on berg deterioration and transport, it is difficult to estimate the degree to which the observed threshold is associated with alternative mechanisms based upon near-surface water mass properties and deterioration suppression by sea ice. Possibilities consistent with the above-described

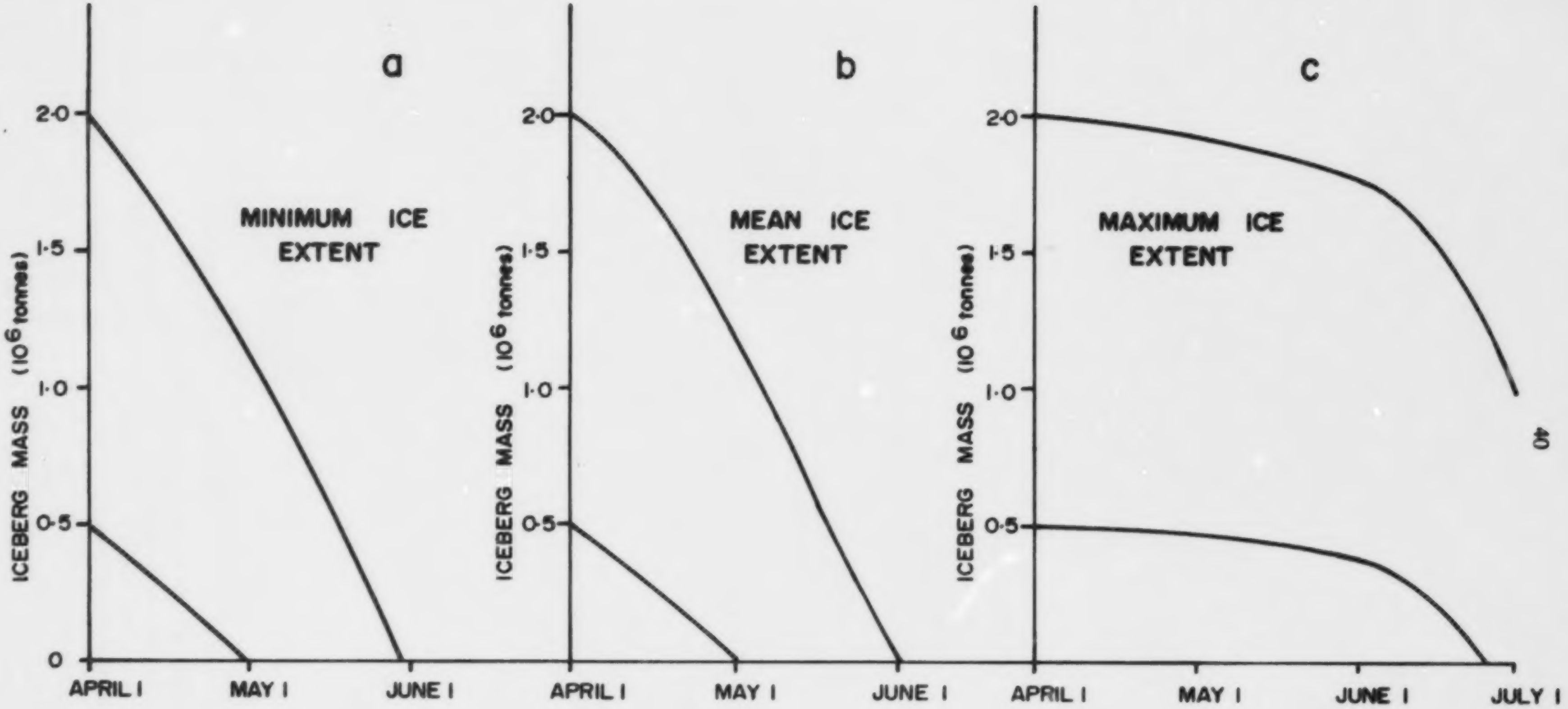


Figure 16 Calculated mass as a function of arrival date at 48°N, 48°W for hypothetical 2.10^6 tonne and $.5.10^6$ tonne bergs under a) minimum; b) mean and c) maximum annual sea ice extent configurations (NOC, 1986).

deterioration -based mechanism could include: reduced probabilities for berg loss to warm offshore waters associated with additional ice and above threshold extents; and the greater (lesser) impact on berg attrition to be expected from a given ice extent increment over the relatively short (long) open water trajectories associated with above- (below-) threshold ice extents.

Because of the limited data and knowledge base, other aspects of deterioration (i.e. large scale berg calving) and variations in other factors (i.e. water temperature, salinity, stratification, sea state and upstream berg fluxes) have been neglected in our simple treatment. Nevertheless, major details of observed past interannual variations in iceberg severities south of 48°N appear to be explicable in terms of the dependence of berg mass loss rates upon the Labrador/Grand Banks spring sea ice extents. The latter events, in turn, appear to be determined, to a considerable degree, through the described combined effects of previous (January) upstream (Davis Strait) ice extents and the directionality and strength of the spring Labrador winds.

The dominant overall role of the sea ice in this picture suggests that this environmental component is, at least initially, the most promising point of focus for consideration of climate change effects on sea ice and iceberg severities in the downstream impact areas of interest.

4. CLIMATE CHANGE EFFECT ASSESSMENT

4.1 Introduction

In light of the above demonstrated dependence of iceberg severity on sea-ice extent, it is sensible for an assessment of possible climate change effects on east coast ice and iceberg regimes to focus on the sea-ice cover. This assessment will include a brief summary of the tools and techniques commonly applied to investigate climate change effects, and a discussion of their adequacy and limitations for the area and period of interest. Difficulties and inconsistencies relevant to sea-ice will be described in some detail (in Section 4.2) to establish the magnitude of the change assessment problem. A major objective of the discussion will be to establish the scope of possible changes in sea ice severities on the basis of present knowledge and to identify feasible steps for providing more definitive judgements. The implication of the sea ice results for iceberg severities are presented in Section 4.3. The latter section also considers, in some detail, contributions to iceberg severity variability which are either independent of, or indirectly related to, the sea ice cover. Again, the scope of possibilities are reviewed and related to available and needed information on tidal glacier dynamics and change-susceptible environmental factors.

4.2 Climate Change and the Regional Sea Ice Cover

The results of Section 3 suggest that the presently observed interannual availability of Labrador/Grand Banks ice conditions may be understood in terms of corresponding variations in earlier mid-winter upstream ice extents and local (downstream) spring wind fields. As a consequence, assessment of the likely effects of future climate change, to a large degree, requires estimation of changes in the January Davis Strait ice extents and the Labrador spring winds. This estimation process inevitably must identify both the mechanisms whereby the latter environmental features are controlled by atmospheric, oceanic and cryospheric parameters and the manners in which these entities are likely to change with continuing modification of the atmosphere.

Global Circulation Models

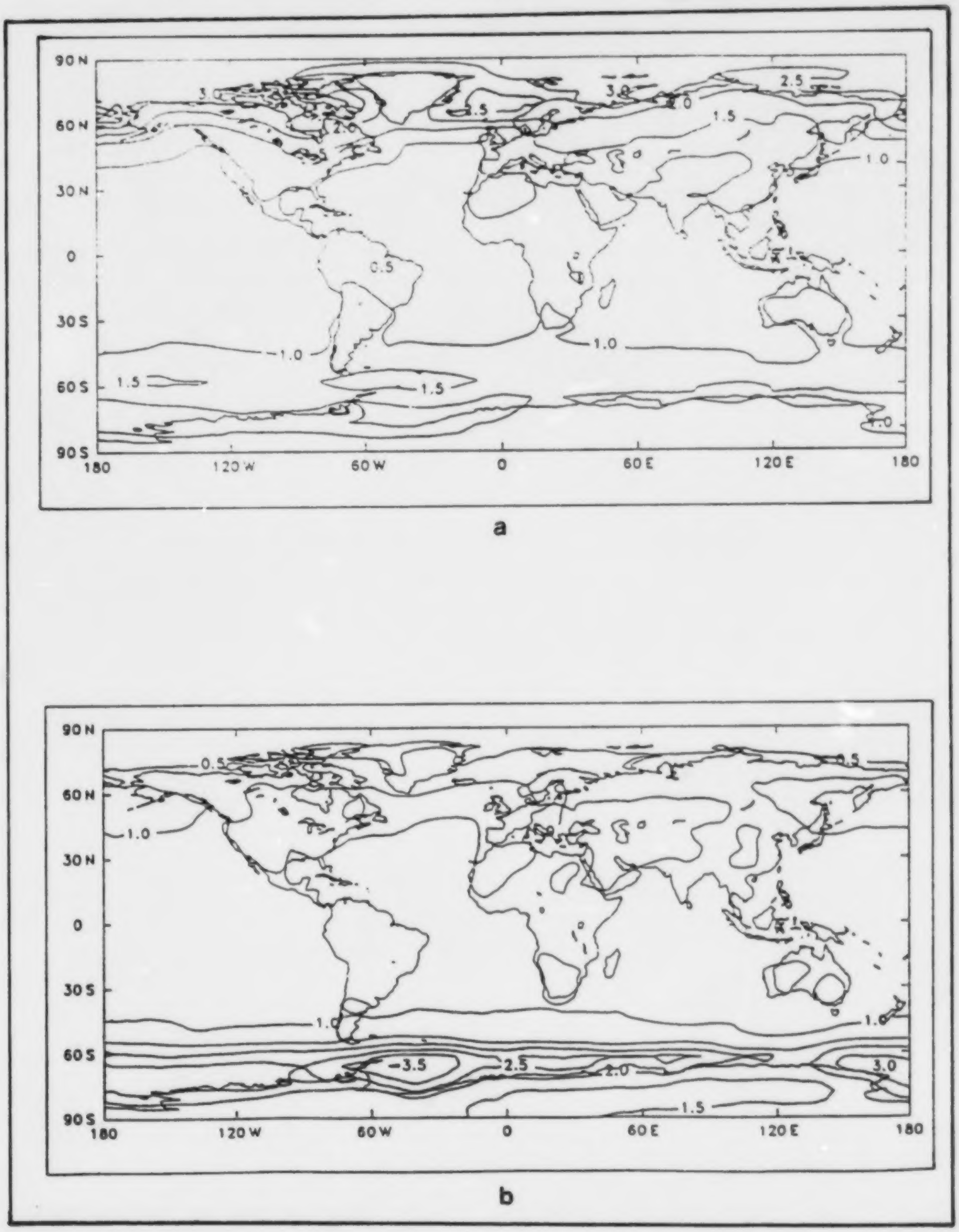
At present, expectations of the extent and pace of anthropogenic global warming are, to a large degree, based upon computer-based models of the atmosphere. The results are obtained through experimental simulations using complex global circulation models (GCMs) run to equilibrium for different assumed concentrations of atmospheric $C0_2$. The most commonly obtained results, equivalent to equilibration of a sudden doubling of present $C0_2$ levels, show strong dependencies upon model construction but, in general, suggest overall global warming will occur. However, significant regional and seasonal differences (Wigley, et al. 1990) occur among various model outputs, which, to some degree, tend to be ignored because of the uncertainties involved in the use of GCMs on the spatial and temporal modelling scales associated with regional and seasonal variability. Thus, for example, present modelling grids (usually several degrees of latitude by several degrees of longitude) do not allow proper representation of ocean currents and eliminate all side channels (Hudson Strait, Lancaster, Jones and Smith Sounds) in our study region. Additional problems are associated with degradation of model performance in areas characterized by high feedbacks such as would be associated with the ice edges and the Greenland ice sheet (Santer and Wigley, 1989). Attempts to deal with such difficulties by, typically, parameterizing sub-grid scale processes, often oversimplify the underlying physics of the environmental interactions.

Ensemble results (Wigley et al. 1990) are often used to reduce the effect of individual modelling errors. Typical ensemble products consist of normalized composites of parameter mapping results obtained from several different model simulations. Although useful, ensemble construction procedures cannot compensate for inaccuracies common to all or most included models.

In general, the results of most modern models run to equilibrium for instantaneous $C0_2$ doubling mirror those of the indicated composite (Figure 17). Warming equal to two to three times the average annual global temperature rise (the quantity plotted in Figure 2) is almost universally indicated for the winter months in a broad strip which encompasses most regions north of 55°N. Typical (Washington and Meehl, 1989) attempts to derive ice boundaries under such warming show significant (5° to 10° latitude) retreats from present day mean ice boundaries although it must be recognized that most models (including that of Washington and Meehl, 1989) overestimate the present winter regional ice extent.

More complicated results, including smaller amounts of atmospheric warming, have also been obtained with a recent "transient" application of the Washington and Meehl (1989) model. In this case, when gradual (1% annually) increases from the present day $C0_2$ level were allowed to occur over a 30 year simulation, modelled winter surface air temperatures were observed to decrease in all but the extreme western portions of our study region. Although the cooling was not statistically significant, it was, in fact, located at the outer limits of a large, statistically significant cooling area centred north of Scandinavia. The apparent sources of this cooling were: (1) a filling and slight northwestward shift of the Icelandic low pressure centre; and (2) weakening of the Atlantic thermohaline circulation (Stouffer et al., 1989). Due to the model's simple thermodynamic ice control mechanism, this cooling allowed a substantial ice edge advance in the GIN Seas region, along with a much smaller, probably statistically insignificant, increase in ice extent adjacent to Labrador.

In view of the apparent disparities among the simulation results and because of the cited inherent limitations of present GCMs, immediate progress toward climate change


90 N
60 N
30 N
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30 S
60 S
90 S
180
120 W
60 W
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120 E
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3.0
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3.0
2.0
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1.0
0.5
1.0
1.0
1.5
1.5
1.0
a
0.5
0.5
1.0
1.0
1.0
-3.5
2.5
2.0
3.0
1.5
b

Figure 17 Contours of enhancements of global-average annual temperature changes anticipated from $C0_2$ doubling on the basis of GCM model composites derived by Wigley et al., (1990) for (a) winter (December-February) and (b) summer (June-August).

assessment is likely to be based upon an initial acceptance of only the most universal GCM result, namely, the elevation of the average annual global surface air temperature. The results of different GCM's vary considerably on regional spatial scales. This large variability reflects the above-noted coarse spatial resolution and crude parameterizations characteristic of the models. As a result, regional and seasonal details, as well as estimates of the rates of change, at present, will have to be obtained through the use of alternative empirical or regional model (sub-model) techniques.

Unfortunately, as indicated in Section 3.2, there are presently no effective regional sub-models with a demonstrated capacity for reproducing the observed interannual variability of regional parameters such as those, the January Davis Strait ice extent and the Labrador spring winds, found to be closely associated with present ice severity variability in the area of interest. As a result, for the moment at least, reliance on the alternative empirical methodology is essential in the assessment process.

Empirical Analogue Techniques

Empirical techniques include:

(1) the identification of regions and times of particular sensitivity to larger scale warming;
(2) the identification of atmospheric configurations which tend to be associated with the extremal conditions of concern; and
(3) the construction of scenarios representative of past responses to global warming.

Results from applications of the first two of these techniques will be given below to both focus our interests and to indicate the types of atmospheric conditions which might accompany drastic changes in the regional ice extent. These results will be followed by a much more extensive utilization of the available data sets to construct both regional scenarios appropriate to global warming and to evaluate and interpret trends in the recorded data sets.

Sensitivity and Relationships between Regional Atmospheric Conditions and Extreme Ice Extent Variations

Sensitivity studies carried out by Cherry (1982) have established the existence of correlations significant at the 95% level between temperatures in a larger scale region (60° to 80°N and 30° to 90°W) and corresponding (simultaneous) ice extents in:

(1) Davis Strait in February; and
(2) the area off Labrador in May.

These results are fully consistent with the linkage between the end of January Davis Strait and Labrador/Grand Bank spring ice extents previously noted in Section 3.2.

The mean atmospheric configurations accompanying extremal ice extents can be computed using a number of different sets of data from groupings of individual or consecutive years associated with alternative high or low values of the January Davis Strait ice index.

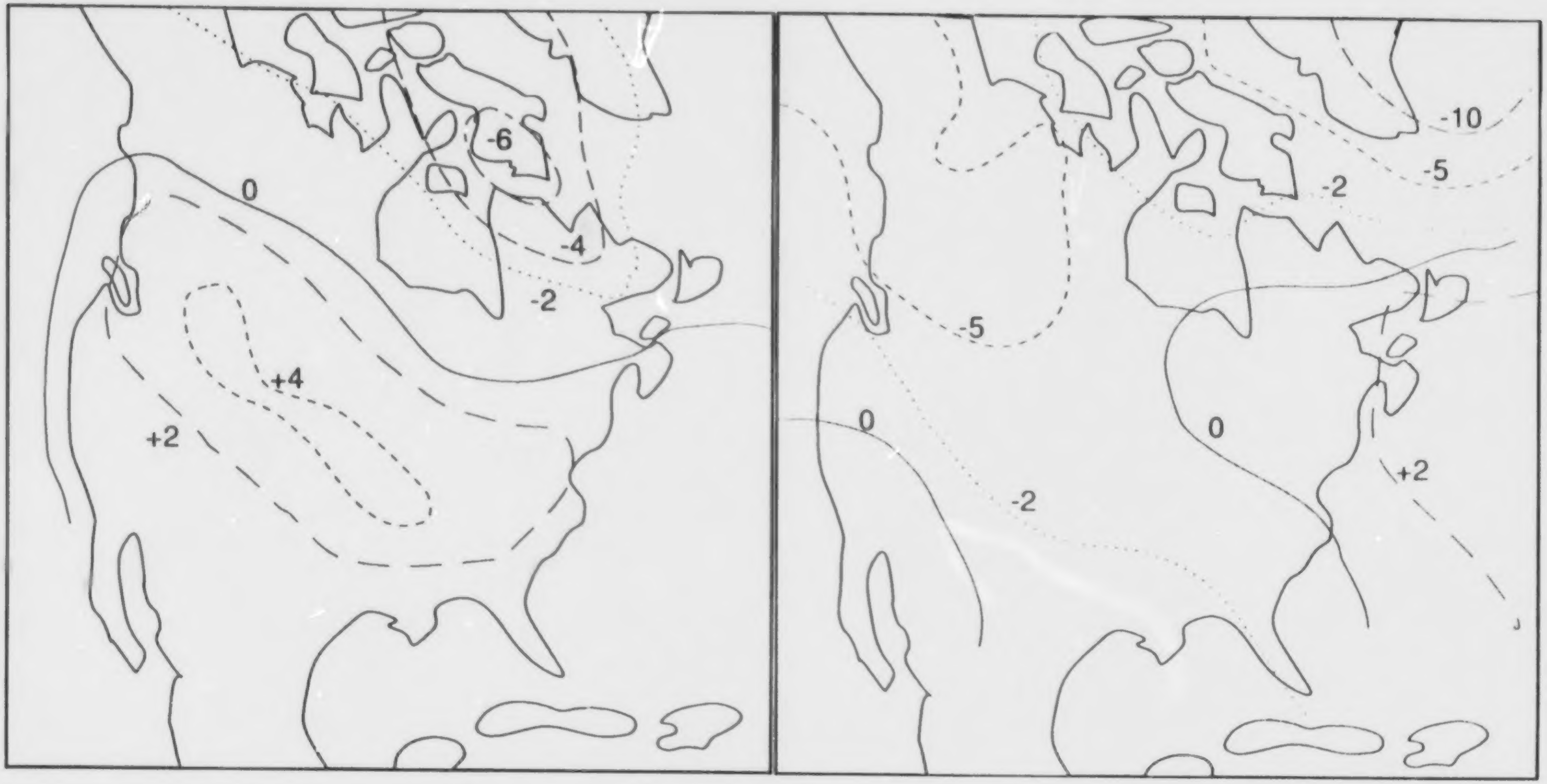
The most dramatic results in this regard were obtained for the mean sea level pressure and temperature difference fields corresponding to two different two-month periods (November-December and January-February) associated (Table I) with the six individually highest values of the latter index (Section 3.2) in the period 1973-1988 and the six individually lowest ice extent years in the somewhat longer period 1961-1988. The extension of the latter period backward in time into the 1960's allowed the inclusion of atmospheric data corresponding to the anomalously low ice extent years associated with that decade (Section 3.2).

TABLE I
Years of High and Low January Davis Strait Ice Extent Used to Derive the Mean Parameter Difference Distributions of Figure 19a, b

HIGH ICE EXTENT	LOW ICE EXTENT
1973	1963
1975	1969
1983	1977
1984	1979
1985	1981
1987	1982

Similar parameter change distributions were noted for both sets of two-month periods, consistent with the previously noted (Walsh and Chapman, 1990), a high persistence of the region's seasonal atmospheric conditions. The magnitude of the differences were, however, larger in the January-February period and the latter results are represented in Figure 19a and 19b. The high ice extent years were found to be accompanied by cooling of as much as .8°C in the study region and by large depressions (as much as 10 to 15 mb) in surface pressure. The centre of the pressure lowering was in the vicinity of East Greenland and Iceland.

Very similar pressure change distribution patterns were derived using the longer consecutive year intervals associated with alternatively high and low ice extents listed in Table II. The nature of the January pressure field differences between the high and low ice extent configurations of the 1980's and the 1960's, respectively, indicates a generally close linkage between regional ice extents and mean winter pressure changes in the vicinity of Greenland and with the mean location of the centre of the Icelandic Low. Comparisons of mean southern Baffin Island-southern Greenland January pressure gradients and Davis Strait ice indices for years in the 1972-1975 and 1983-1987 high ice extent period showed the existence of a good linear relationship (r=.92) between the two quantities (R. Brown, personal comm.).



January - February Temperature Difference (°C)   January - February Pressure Difference (mb)

Figure 19 Average (a) temperature and (b) sea level pressure differences between mean January conditions in six maximally high ice extent minus six maximally low ice extent years (see Table I) using ice extents measured in Davis Strait at the end of January.

TABLE II
Pairings of Low and High Davis Strait Ice Extent Periods Used for Scenario Constructions (Ice Estimates Based on Environment Canada and U.S. Naval Oceanographic Centre Charts and Walsh, 1978)

High	Low
1983-1987	1976-1980
1972-1975	1960-1963
1980-1988	1960-1969

The indicated correlation between large January ice extents, a deepening of the Icelandic Low and lower regional winter temperatures, is consistent with the known details of the ice cover forcing and the effects of pack ice on atmospheric temperature. Clearly, the deepening of the Icelandic Low pressure centre increases the east-west pressure gradient, and hence the northerly winter wind components in the study region which were previously found (Section 3.2) to be associated with high Davis Strait ice extents. Likewise, an accompanying regional cooling would be expected because of the resulting decreased flow of sensible and latent heat between the ocean's surface and the atmosphere and because of the elevated regional albedo.

Although the presence of feedback complicates precise definitions of cause and effect, such associations between past atmospheric variations and extreme Davis Strait ice extents provide useful guides in assessing the implications of future atmospheric changes anticipated on the basis of climate model and empirical scenario results.

Empirical Global Warming Scenarios

The scenario approach for identifying responses to global warming is based upon the assumption that, given approximately similar boundary conditions (i.e. ocean temperatures, ice extents, etc.), future changes in the regional climate will be relatively independent of the source of the global perturbation. As a consequence, in this view, warm periods from the past can be used to provide information on the potential regional impacts of global warming through the assumed similarity of the effects of such warming on atmospheric circulation and temperature. The requirement for similar boundary conditions restricts, at best, the applicability of the constructed scenarios to periods of duration shorter than those required for major adjustments in atmospheric, oceanic and ice environmental structures.

Scenarios are constructed in the form of mappings of sea level temperature and pressure differences associated with groupings of years characterized by alternatively high and low surface temperatures in the Arctic as a whole, as well as on hemispheric and global bases. These parameter field differences have been computed for groupings composed of both isolated individual warm (cold) years and blocks of 5, 10 or more successive years of, on average, warmer (colder) temperatures (Palutikof et al., 1984).

Selections of warm and cold years for scenario construction are usually based upon the recent compilations (Hansen and Lebedeff, 1987; Jones et al., 1987) of worldwide temperature data recorded since the late 19th century. The computed global and northern hemispherical mean temperatures (Figure 18) show a relatively steady (on a decadal

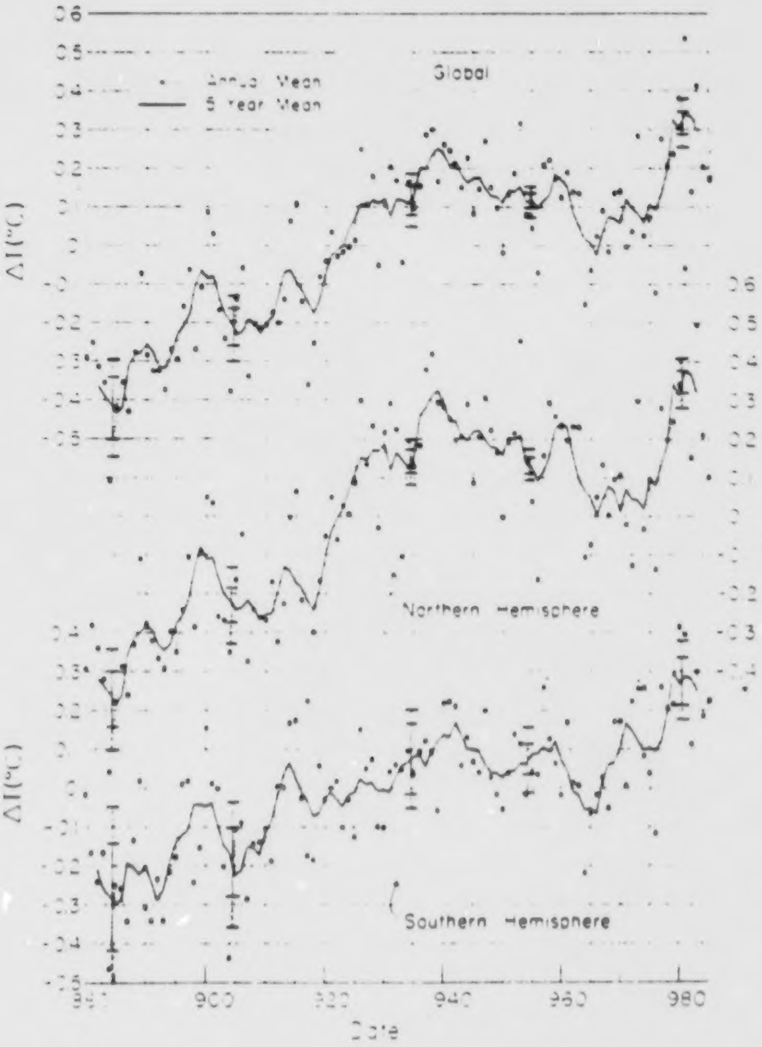


Figure 18 Global and hemispheric surface temperature changes as estimated by Hansen and Lebedeff (1987) from meteorological station records. The 5 year running mean is a linear average centred on the plotted year with inner and outer error bars referring to the 5 year- and annual- means respectively.

timescale) warming of .5° C between 1880 and 1940; a cooling of about .2°C between 1940 and 1970, and a subsequent warming to the present day of approximately .4°C.

A potentially significant difference between the two noted warming periods was observed in the zonal dependence of temperature change. In the earlier (1880-1940) period, the observed warming was largely a northern hemisphere effect, strongly increasing with latitude (the average warming in the Arctic was approximately 2°C as opposed to perhaps .3°C in the 0 to 24°N zone). A similar latitude dependence was evident in the subsequent cooling which was largely associated with sharply lower Arctic temperatures in the period 1940-72. The on-going warming of the 1970 and 1980 decades, on the other hand, was found to be much less dependent upon latitude, for, at least, areas poleward of roughly 45°.

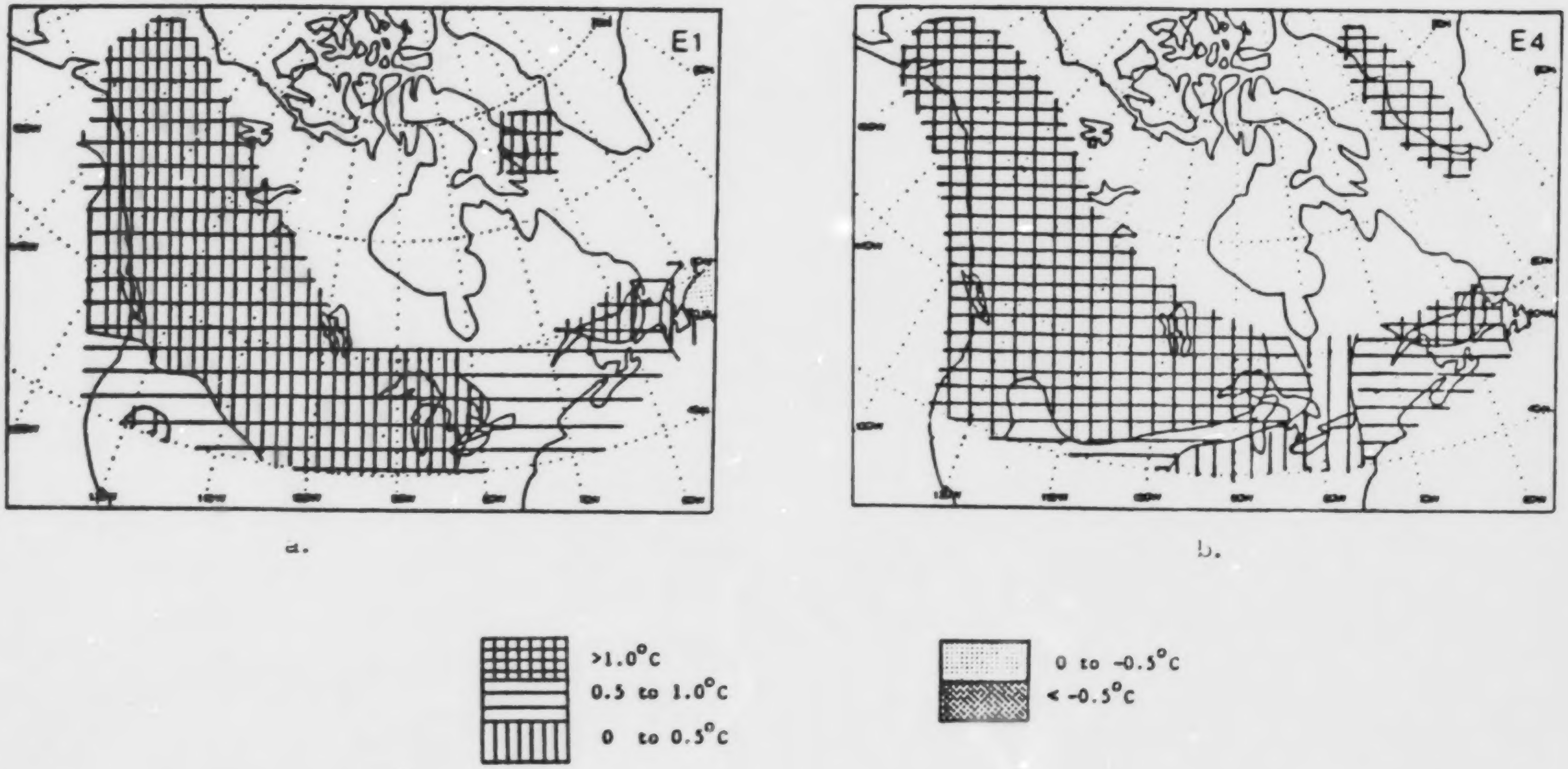
Unfortunately, empirical scenarios constructed on the basis of the indicated global/hemispheric temperature trends have not shown the consistency which should be associated with a single source-independent regional climate response to larger-scale warming. I noticed at least two different responses were noted.

One of these responses corresponded to higher winter temperatures throughout all but, in some cases, the southernmost portion of the study region (Figures 20a, b). The accompanying (Figures 20c, 20d) characteristic pressure changes were similarly positive except for some decreases in the extreme north and to the west of Labrador and Baffin Island. Although the scenarios of Figure 20 were derived with relatively sparse and lower quality pre-1924 data, similar patterns were associated with pairings of more recent time interval data. The January mean pressure differences for one such pairing, specifically for the 1979 to 1983 and 1968 to 1972 warm and cold periods, respectively (Figure 21), corresponded to an overall difference in the mean global temperatures of the two intervals equal to about .3°C (Hansen and Lebedeff, 1987). The more detailed contouring in the latter figure shows a slight extension of the northern pressure depression (relative to the scenarios of Figure 20), but an overall warming response which is equivalent to a filling of the Icelandic Low.

The second type of response was apparent in several different scenario constructions, including a pairing of the 1968 to 1972 cold interval (previously used in the derivation of Figure 21), and a 1944-1948 warm interval (Figure 22). In this scenario type the warmed winter atmosphere shows a very widespread depression of pressure in the study region frequently, with maximum decreases centred east of southern Greenland. A similar characteristic change in winter pressure was also noted in the seasonal patterns constructed from five isolated maximally warm years and five isolated maximally cold years (Figures 23 and 24). The representivity of the temperature scenario results was limited by the restriction to land station data and by the particularly sparse data sets available for Baffin Bay and Davis Strait.

Nevertheless, comparisons show that in this response mode, regional warming is confined to the spring and summer periods, with the most pronounced effects of the larger scale warming being fall and winter cooling throughout all or most of the study region coincident with the winter intensification (deepening) of the Icelandic low pressure centre.

As indicated above, because of the cited differences, the body of scenario results does not support the obvious existence of a single consistent regional response to larger-scale warming. Difficulties in this regard are most apparent in the very different pressure difference fields of Figures 21 and 22 which were derived using a common (1968-1972) cold interval, and two different (1944-1948 and 1979-1983) warm intervals. The evident


E1
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a.
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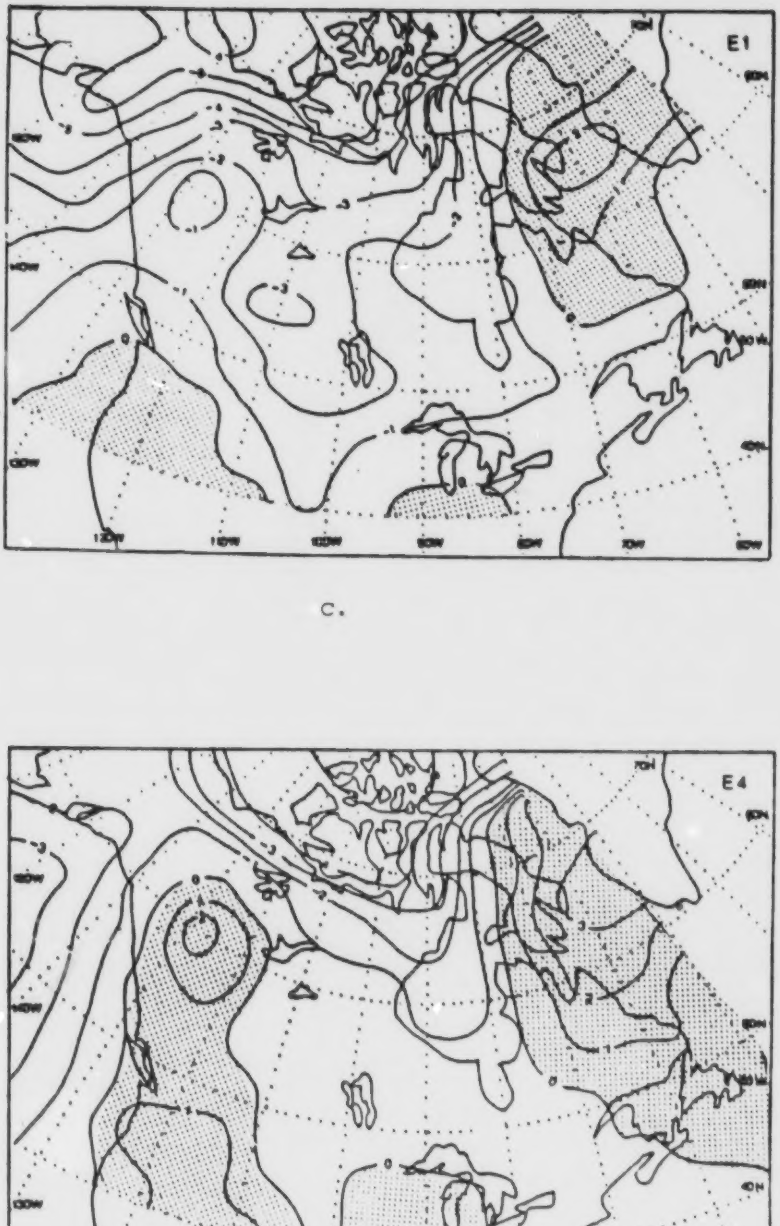
>1.0°C
0.5 to 1.0°C
0 to 0.5°C
0 to -0.5°C
< -0.5°C

Figure 20 Winter temperature change scenarios: (a) surface temperatures 1944 to 1948 (warm minus 1916 to 1920 (cold); (b) surface temperatures 1959 to 1963 (warm) minus 1916 to 1920 (cold); continued on next page...


E1
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60W
c.
E4
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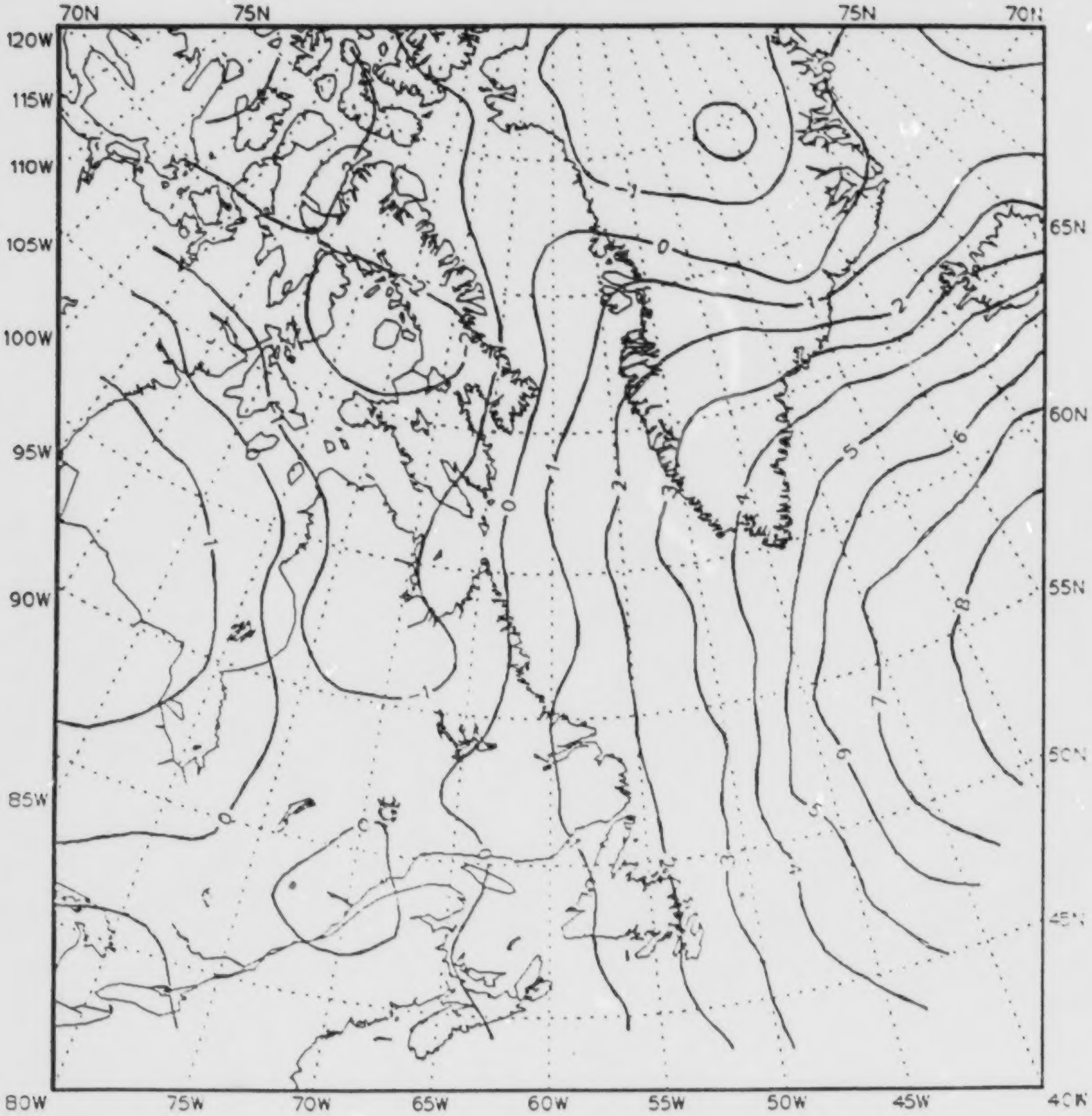
70W
60W
d.

Figure 20 Winter pressure change scenarios: (c) sea level pressures 1944 to 1948 (warm) minus 1916 to 1920 (cold); (d) sea level pressures 1959 to 1963 (warm) minus 1916 to 1920 (cold) [from Brown and Walsh, 1986].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0
1
2
3
4
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6
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8
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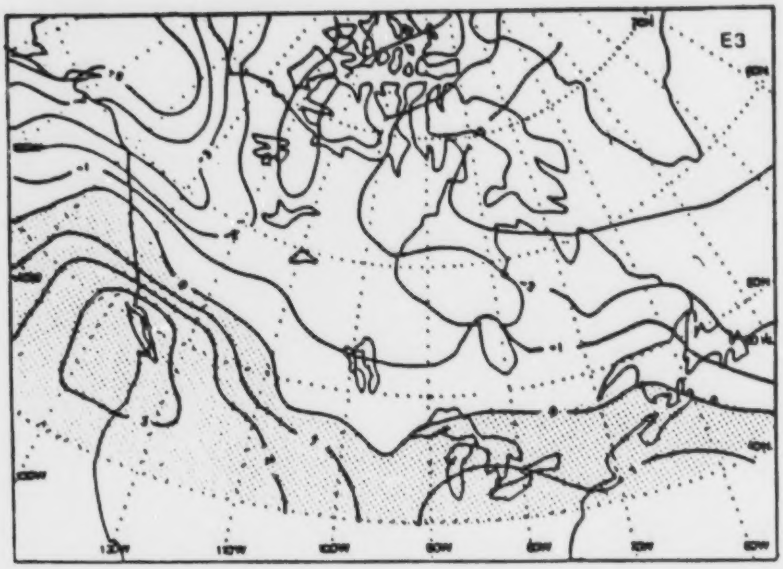
Figure 21 Winter sea level pressure change scenario for January 1979 to 1983 (warm) minus January 1968 to 1972 (cold) years [calculated from Environment Canada data by R.D. Brown (pers. comm.)].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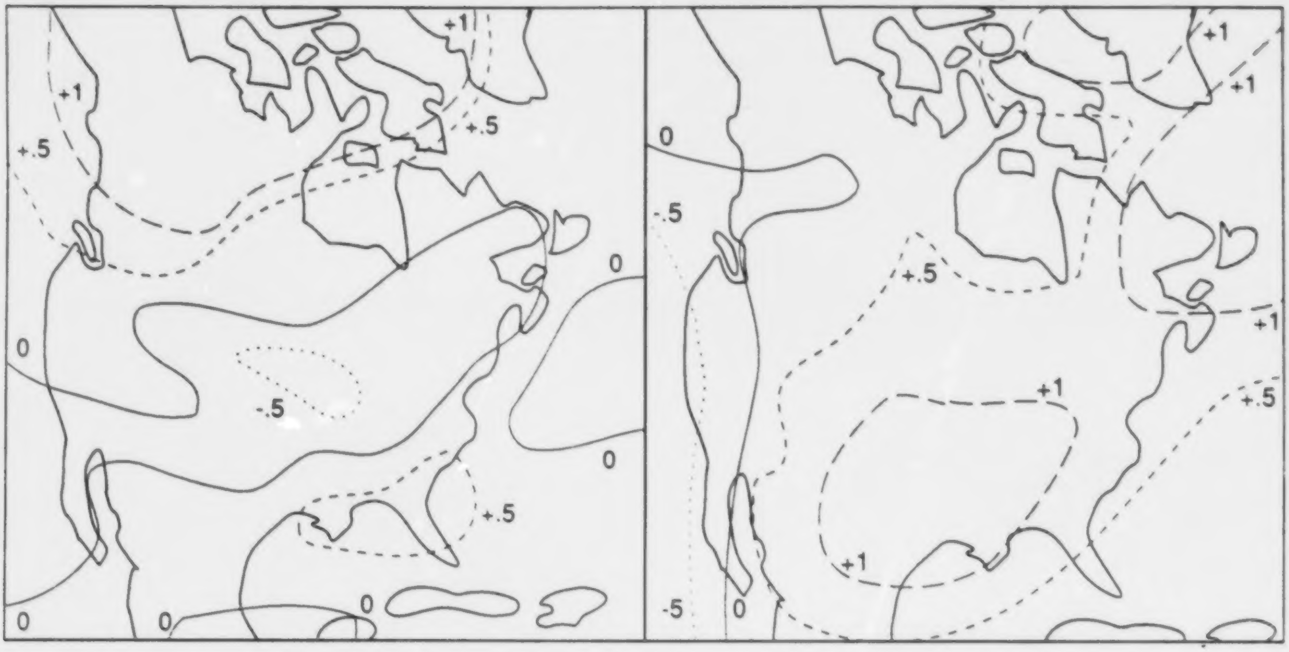
70W
60W

Figure 22 Winter sea level pressure change scenario for 1944 to 1948 (warm) minus 1968 to 1972 (cold) [from Brown and Walsh, 1986].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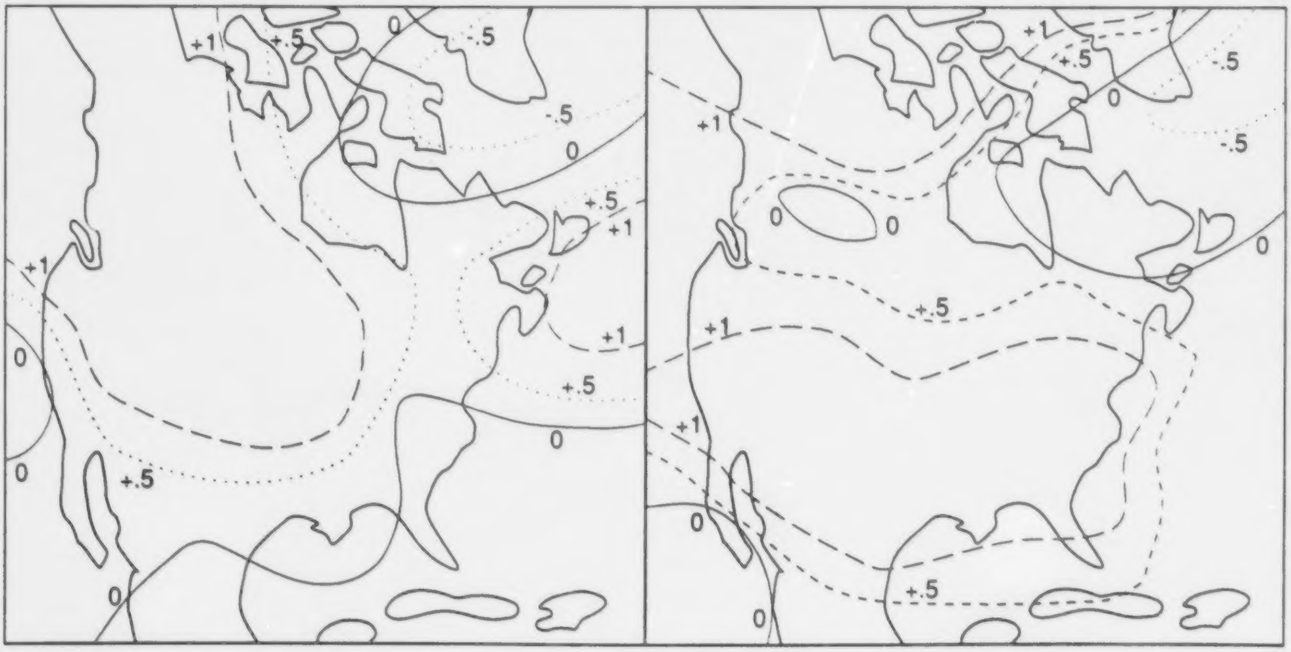
Spring

Summer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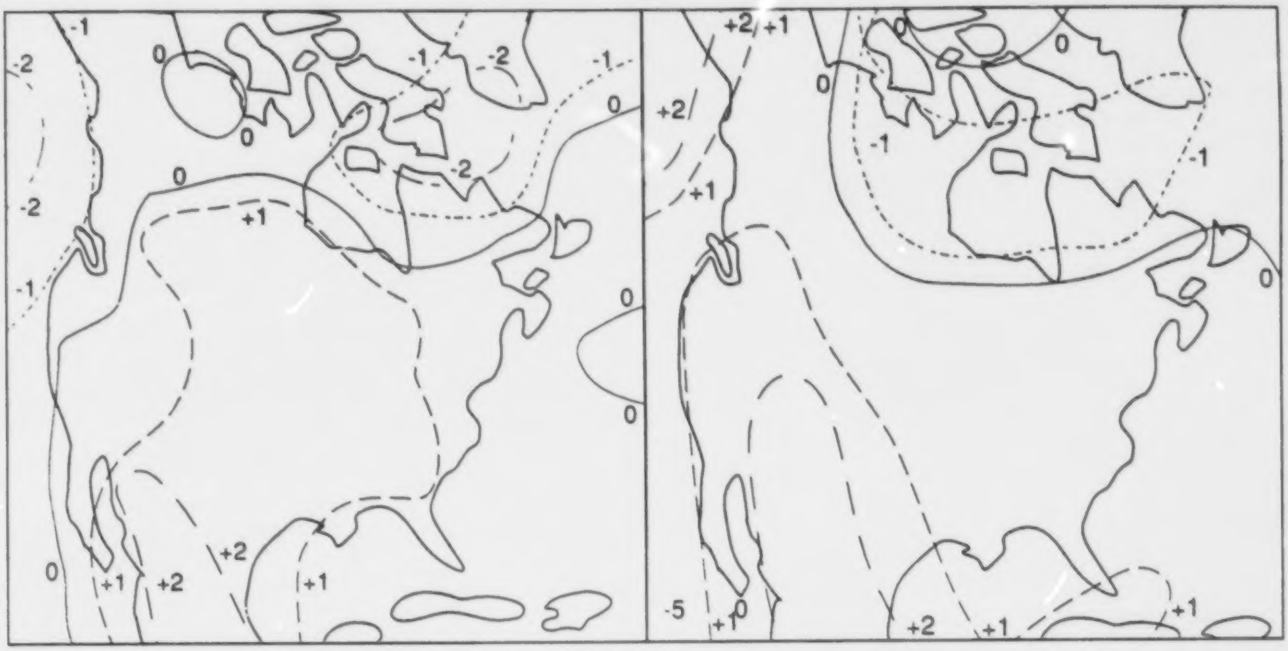
.5
+1
+.5
0
-.5
-.5
0
0
0
+1
+.5
+1
+1
0
+.5
0

Autumn

Winter

Figure 23 Temperature difference contours relative to a group of individually warm-minus a group of individually cold-years (Palutikof et al, 1984). The warm years in this scenario were 1933, 1937, 1938, 1943 and 1953, while the cold years were 1955, 1964, 1966, 1968 and 1972.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+1
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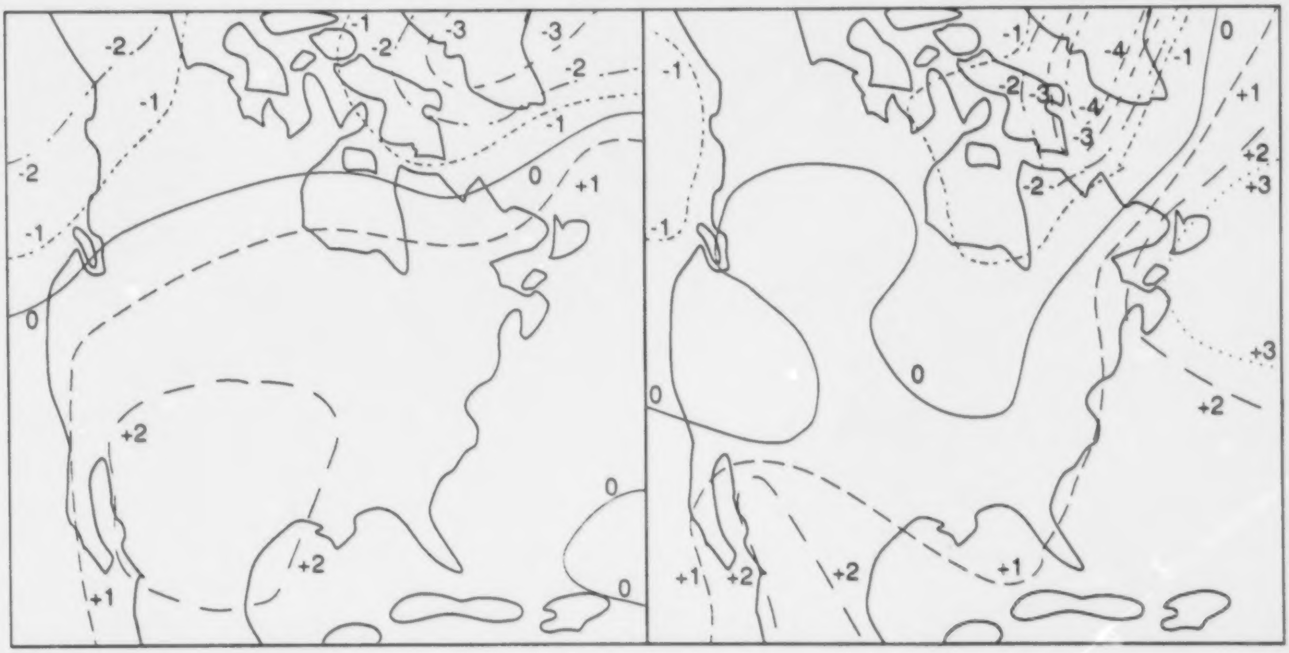
Spring

Summer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-1
-1
-4
-1
0
-2
-3
-4
-3
+1
-2
+2
+3
-1
0
+3
0
+2
+1
+2
+2
+1

Autumn

Winter

Figure 24 Seasonal sea level pressure change scenarios from 5 maximally warm minus five maximally cold years (see Figure 23) [from Palutikof et al., 1984].

dependence of the sign of the mean pressure changes on the choice of the warm interval suggests that empirical scenarios constructed for intervals even as long as 5 or 10 years do not give unambiguous estimates of likely changes in regional atmospheric parameters corresponding to simultaneous larger scale (global/hemispheric) temperature change. Unfortunately, it is not presently possible to circumvent this difficulty by following recommendations (Palutikof et al., 1984) that comparisons be based upon data from longer (20 years) comparison intervals. Problems in this regard arise from the absence of large differences in the global mean temperatures (Figure 18) associated with different 20 consecutive-year periods in the post-1924 period coincident with reasonably reliable temperature data.

It should be noted, at this point, that the general patterns previously noted to correspond with high Davis Strait ice indices tend to resemble, most closely, those associated with the second discussed general type of regional response to global warming. Regional pressures decrease in both instances with greatest depressions usually present east of Greenland and significant winter cooling occurs in the region. In the global warming scenario (Figure 22), the cooling is confined to the southern portion of the region as opposed to the more generalized extension of lower temperatures under high ice extents. Nevertheless, it is not possible to make any assertion of simple correspondences between the atmospheric configurations associated with, respectively, globally warm and high regional ice extent periods. For example, the years 1972 to 1975 used to derive high ice extent atmospheric configurations (similar to Figure 22) relative to 1976-1980 (Table II), were actually associated with a mean global temperature which was lower than that of the latter period.

As well, there was no evidence of definitive global warming-induced changes in the southern Baffin Island-southern Greenland sea level pressure gradient similar to those noted above with respect to Davis Strait ice extents.

Trends and Implications of Empirical Results

Under these circumstances, it would appear that further applications of the empirical methodology to the climate change assessment problem would have to be limited to the identification of longer term trends in the interrelationships of global temperature, regional sea ice extents and the regional atmospheric parameters. By focusing on changes occurring on temporal scales no shorter than decadal, this approach could, at least in principle, reduce the "noise" in the climate-change signal contributed by poorly understood quasi-periodic phenomena such as the North Atlantic Oscillation (NAO), and by teleconnections to the El Nino Southern Oscillation (ENSO).

Unfortunately, the application of this approach to the regional problem is limited by the very small amount of surface temperature and pressure data available for areas north of Labrador prior to 1945, and by the dearth of reliable sea ice extent data for Davis Strait and off Labrador/Grand Banks prior, respectively, to 1972 and roughly 1950.

Recent analyses of temperature data (Hansen and Lebedeff, 1987; Jones, 1988) have indicated the occurrence of a distinct cooling trend in the study region over the period 1965-1985. This decrease in the regional mean atmospheric temperature runs counter to the larger scale hemispheric/global warming observed over the same period. Such an out-of-phase regional temperature response was not identified in the earlier (1880-1940 and 1940-1965) periods when regional temperature changes of +3° to 5°C and 0° to -1°C, respectively, were similar in sense to those observed on global/hemispheric scales. Further examination of this response was recently carried out (R.D. Brown, pers. comm.) through comparison of a composite of two Newfoundland/Southern Labrador temperature records (data recorded at Belle Isle prior to 1969 and subsequently at Cartwright) with the mean northern hemispheric record (Jones et al., 1987) (Figure 25). Aside from a lack of a detectable response to a 1935-1945 hemispheric warming, the regional temperatures tended to follow the hemispheric trend until, roughly 1960, when the subsequent out-of-phase response pattern became apparent.

On the other hand, as indicated above, fully reliable data on the important January Davis Strait ice index only became available in 1972, coinciding with the approximate beginning of the second global warming episode (Figure 18). A plot of these data (Figure 26) along with ice extents derived from the Walsh (1978) compilation for the years 1953 to 1971 shows a very distinct tendency toward higher ice extents over the past two decades. For example, the 1.0 index value previously associated with expectations of high berg severities (Marko et al., 1986) has been exceeded in 7 of the last 9 years after only 11 such exceedences in the previous 30 years. The mean indices for the 1953-60, 1961-70, 1971-80, and 1981-90 intervals were 0.84, 0.88, 0.97 and 1.2, respectively.

Again, some caution must be used in interpreting these results in view of the uncertainties in the Walsh data set used to derive the pre-1972 ice extent values. The data from the 1953 to 1960 period are a particular problem with from 45 to 100% of the measurement region data points in individual years being denoted as "estimated" because of: discrepancies between equivalent data sets; specific notations of uncertainty by observers; or absence of reported data points.

In order to both provide other evidence of ice extents in the pre-satellite years and, as well, to extend the data base on ice extent variability further back in time, a number of additional sources of data were reviewed. These data sources were all indirect with respect to the Davis Strait ice index and were based upon correlations demonstrated in this study relative to the latter index. Our use of these "proxy" data sources was intended to provide multiple, independent indicators of decadal trends in order to minimize, as much as possible, misjudgments based upon errors in any given data set.

Because of the very different measurement sensitivities and accuracies corresponding to different decades in the period of interest, emphasis in proxy selection was given to identifying the occurrence frequency of extreme ice conditions. This approach is sensitive to thresholding but, in the absence of a better alternative, appeared to give reasonably consistent results. The individual proxies considered are described in the following paragraphs.

1. Iceberg Counts - The empirical relationship between Davis Strait indices $I \geq 1.0$ and iceberg counts $N \geq 600$ proposed by Marko et al. (1986) was assumed to have been valid throughout the period of the historical iceberg count record. By ignoring the changes in reduced counting efficiency and lesser accuracy throughout the earlier period, this assumption allowed an estimate of the minimum number of high ($I \geq 1.0$) indices per decade. (Spring Labrador winds could have reduced berg counts below the high severity threshold in spite of preceding high ice indices.)

2. S. Labrador/Newfoundland Ice Index - The empirical relationship between the Davis Strait ice index and the spring ice extent south of 55°N can be used to identify trends in I. For this purpose, we use the recent results compiled by Hill and Jones (1990) for ice area south of 55°N in March. Based on an empirical comparison with the Davis Strait Ice Index from 1972 to 1988, March ice extents exceeding 2.5×10^5 km^2 from

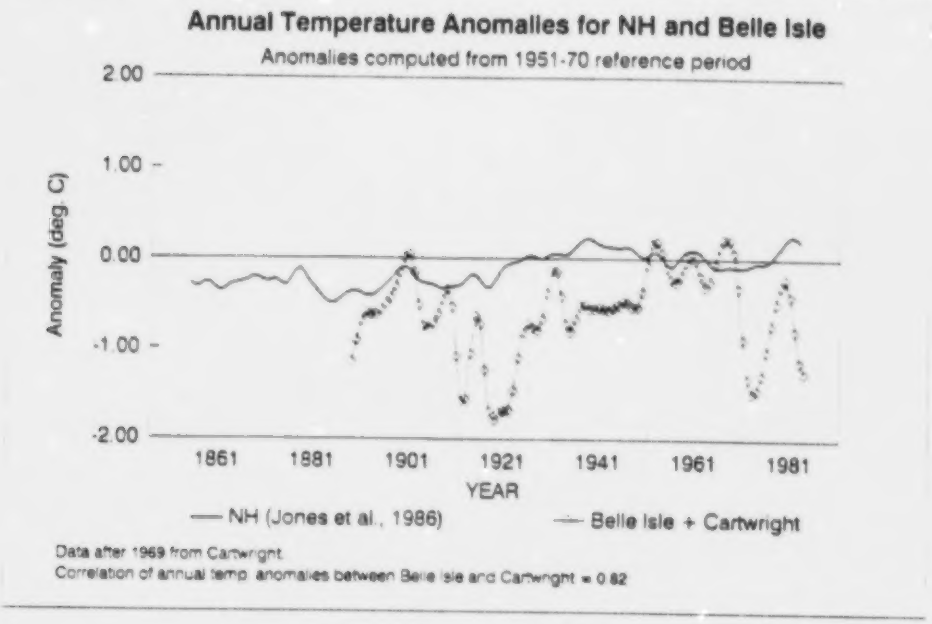


Figure 25 A comparison of the composite annual temperature anomaly as computed at Belle Island - Cartwright with the global mean temperature anomaly as computed by Jones et al., 1986 (R.D. Brown, pers. comm.). The combination of Cartwright and Belle Isle data was justified on the basis of the fact that the former station was both the closest monitoring site to Belle Isle and correlated most strongly (r=0.82) in the common period of overlap (1935 - 1969). Further, no discontinuity was evident in the plot of the temperature anomaly differences. The mean value of these differences for 1935 - 1969 was 0.07, lower than any pairing of other stations with Belle Isle in their respective overlap periods. Both time series were smoothed with a 9-term binomial filter.

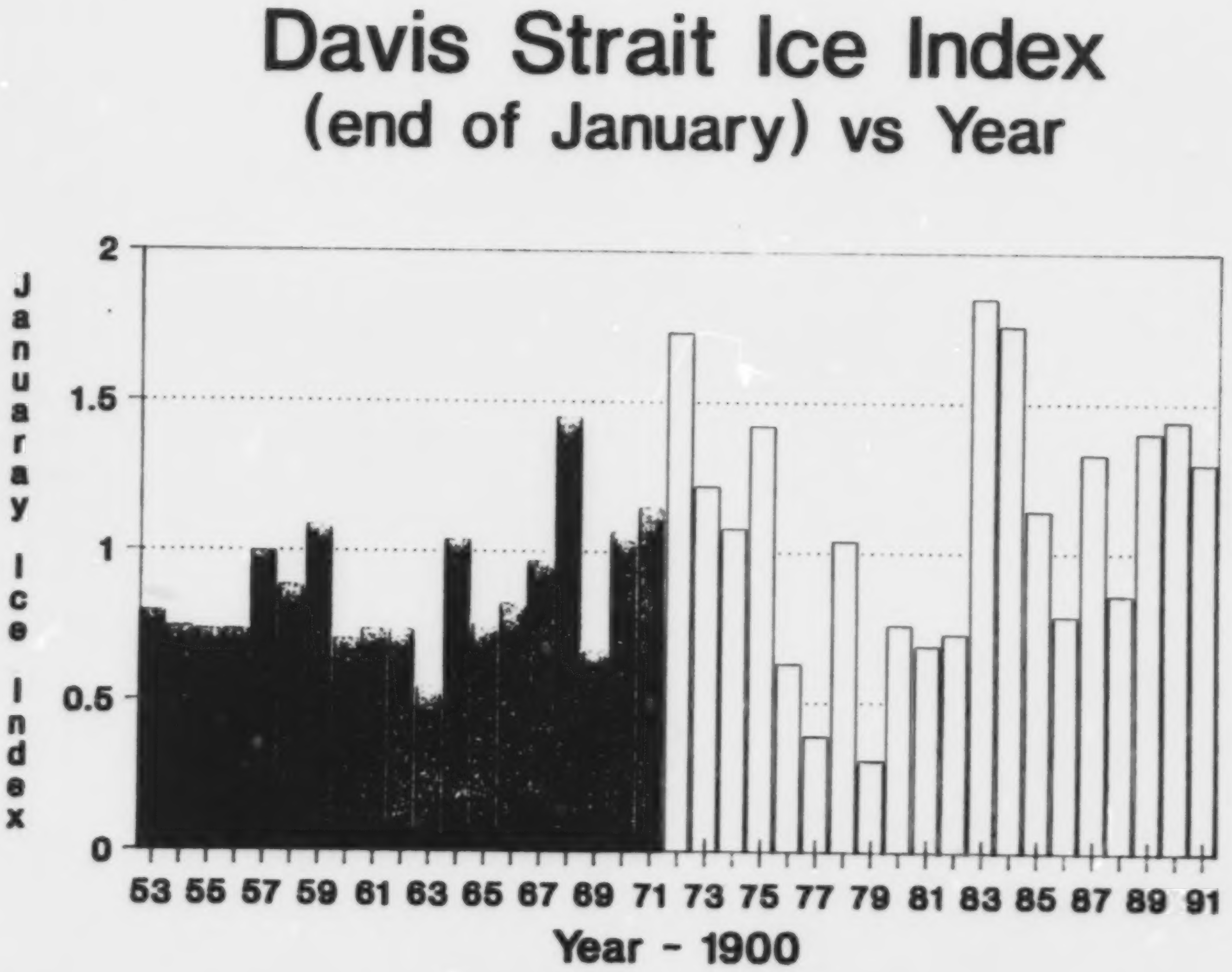


Figure 26 Sea ice extent in Davis Strait in late January from 1953 to 1991. The sea ice extent for the period (1953-1971) prior to availability of routine satellite measurement was derived from the Walsh (1978) data set.

the Hill and Jones (1990) data set were identified as corresponding to high Newfoundland ice extents. Again, because of possible occurrences of easterly spring winds, trends in this index are representative of trends in the minimum number of high Davis Strait ice indices per decade.

3. Frobisher Bay (Iqaliut) Air Temperature - Extension of results obtained in an earlier study (Marko et al., 1986) to cover the period 1972-1991 has shown that, with one exception, all years with mean January Frobisher Bay surface temperatures below -28.2° were associated with high January Davis Strait ice indices. The one exception to this rule, 1985, was also the single exceptional year in which a high index value (see above) was not coincident with strong northerly wind components in Davis Strait. Because of this exception, one can assume that this proxy also provides a lower limit estimator of the decadal numbers of high Davis Strait extent values.

4. Station 27 Coldwater Signatures - During the study, it was noted that with one or possibly two exceptions, during the 1972-1989 period of good Davis Strait ice extent data and, as well, during the two known instances of high berg severities during the 1946-1971 period, occurrences of high January Davis Strait ice indices were coincident with continuous negative surface water temperature anomalies during February through April at a monitoring station off the Newfoundland coast (47° 33'°N 52° 35'°W) (Keeley, 1981). In the one known exception (1983), negative anomalies following the occurrence of a high annual Davis Strait ice index, were confined to the water layers below 30 m. As a consequence, trends in occurrences of this signature, again, represent estimates of trends in minimum numbers of high Davis Strait index occurrences per decade.

The decadal occurrences of high values in each of these Davis Strait ice extent index proxies are listed in Table III along with the actually observed occurrences for the two most recent decades and the high value occurrences as deduced from the Walsh (1978) data compilations for the period prior to 1972. Also included in Table IV are the previously noted mean values deduced both from the post satellite data and the earlier Walsh (1978) data set.

The general trend of the Davis Strait ice index (as could be inferred from Figures 26, 12c, Table III and the above-cited means for intervals in the 1953 - 90 period) appears to be consistent with the occurrence of a pronounced minimum in the regional winter and spring ice extents during the 1950 and 1960 decades. On the basis of the probably more reliable downstream ice extent proxies (see also Figure 12c) and the Iqaliut temperature data, it seems most likely that the actual minimum in the upstream ice extents occurred during the 1960's.

This trend both between the 1950 and 1960 decades, and relative to earlier and later decades, is also evident in the decadally averaged berg counts (Table IV). These averaged quantities showed less than 10% variation over the period 1901-1950. The noted decline in the subsequent two decades is hard to explain solely as a measurement artifact in view of the improvements in counting technologies (unless earlier results were heavily contaminated with multiple recounting of individual bergs). Again, although the introduction of SLAR into the counting process in 1983 complicated comparisons, similar ratios are obtained between average annual values of the berg counts and ice extents for both the 1971-1980 and 1981-1989 periods. In other words, the mean berg counts scaled with mean ice extents for, at least, the two best studied decades.

TABLE III

January Davis Strait Ice Index Means and Numbers of high annual values by decade. Decadal high value occurrences given in years in Columns A-F for both observed/recorded values of Davis Strait index and for each of the proxy indices defined in the text. (Numbers in brackets indicate, for incomplete decade data sets, numbers of years of available data)

DECADE	MEAN INDEX	OBSERVED HIGH DAVIS ST. INDICES A	HIGH DAVIS ST. INDICES (WALSH DATA) B	HIGH PROXY #1 C	HIGH PROXY #2 D	HIGH PROXY #3 E	HIGH PROXY #4 F
1921-30				2	4		
1931-40				3	3		
1941-50				3	4	3(5)	1(5)
1951-60	.84		2	2	4	4	2
1961-70	.88		3	0	0	1	1
1971-80	.97	5(9)	6	3	5	5	4(7)
1981-90	1.2	6	6	3(9)	4(8)	5	4(8)

TABLE IV

Average Annual Iceberg Counts South of 48°N as Estimated by the International Ice Patrol

DECADE	AVERAGE ANNUAL COUNT
1901-1910	425
1911-1920	410
1921-1930	462
1931-1940	421
1941-1950	414
1951-1960	236
1961-1970	155
1971-1980	439
1981-1989 (9 yrs)	519

Taken as a whole, these data are consistent with a general decrease in regional winter and spring ice extents in the last two decades of the nominal 1940 - 1970 period of global cooling (Hansen and Lebedeff, 1987). This decline was most intense in the 1960 decade associated with both the most prominent temperature decrease of the period and the start of the out-of-phase (relative to hemispheric conditions) response of the regional temperature (Figure 25). The increased winter ice extents of the subsequent two decades correspond with expectations in terms of the accompanying deepening of the Icelandic Low pressure centre and the winter and overall annual regional cooling characteristic of these periods (Hansen and Lebedeff, 1987; Jones et al., 1987).

A conservative interpretation of the available air temperature and sea ice databases is that, short of an unexplained decline associated with the latter half of the 1940-1970 global cooling episode, there is no evidence of a decrease in regional sea ice severities over the years between 1900 and 1990. During this period, atmospheric levels of CO_2 increased by more than 20% and global temperatures rose by about 0.7°C. On the contrary, over the last two decades, the mid-winter ice extents in Davis Strait have been increasing on average by approximately 2% annually, coincident with regional cooling of over 0.5°C. This behaviour also occurred during a period when estimates for the Arctic as a whole (Mysak and Manak, 1989) have shown a decline in sea ice extent. It may also be significant to note that the shift from in-phase to out-of-phase relationships between the regional and hemispheric/global temperatures (Figure 25) which occurred at the beginning of the anomalous 1960 decade closely resembles a similar shift in the phase relationships between the ice extents off Labrador/Grand Banks relative to those in the GIN seas region (Hill and Jones, 1990). The mechanism proposed for the latter shift (Hill and Jones, 1990), namely a positive pressure anomaly centred on northern Greenland, could explain the relatively low and high ice extents observed in our study region and in the GIN Seas, respectively, during the 1960 decade. On the other hand, the out-of-phase relationship has continued through the subsequent decades which have been associated with an intensification of the Icelandic Low pressure centre.

It should be noted that the anomalous 1960 decade was coincident with a major decrease in the atmospheric climate forcing as calculated by Hansen et al. (1984) and Hansen and Lacis (1990) using simple models (Figure 27). As indicated in the Figure, the reduced atmospheric forcing during the 1960's was associated with volcanic eruptions which produced temporary increases in stratospheric aerosol concentrations. Detailed comparisons of Figures 18 and 27 show, with the possible exceptions of a brief period centred on 1914, that intervals of decreased forcing have corresponded fairly closely with local minima in the smoothed temperature records. One such decrease was associated with the first half of the 1980 decade. The resulting cooling was, however, of distinctly shorter duration relative to that observed two decades previously, perhaps reflecting the larger influence of the greenhouse forcing under continuing atmospheric modification.

Quantitative anticipation of the details of future global warming will almost certainly require separate identification and characterization of other shorter term and oscillatory climatic perturbations associated with volcanic and other phenomena (such as solar cycle variations, Hansen et al., 1984; Hansen and Lacis, 1990b; and NAO and ENSO cycles, Walsh et al., 1986; Marko et al., 1986; Myers et al., 1988).

Unfortunately, the lengths of the temperature and sea ice data records are not sufficient to allow judgement of the extent to which the regional changes observed since roughly 1960 are representative of global warming effects as opposed to such unrelated periodic or aperiodic phenomena. Other types of empirical data have contributed little clarification

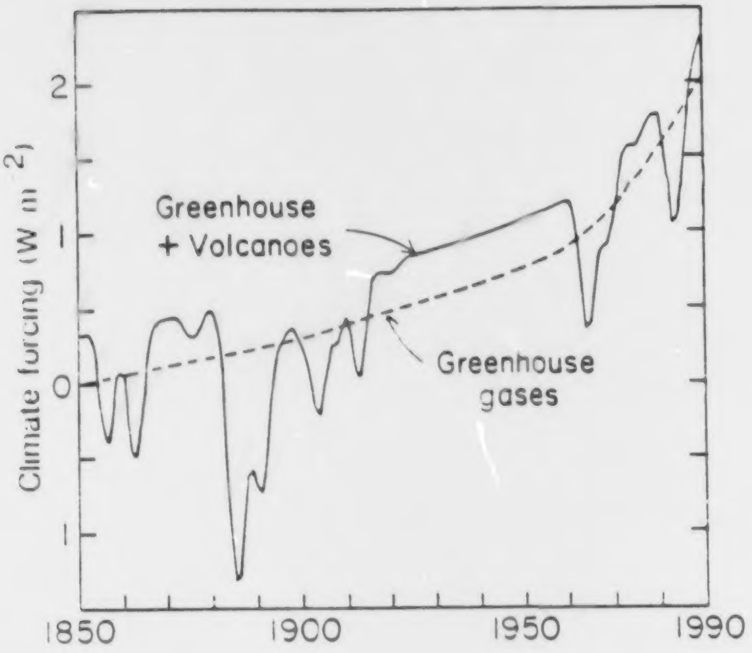


Figure 27 Estimated climate forcing in the twentieth century under greenhouse gas concentration change and observed changes in stratospheric aerosols due to volcanic activity (Hansen and Lacis, 1990).

in this regard. Thus, recent declines in overall Arctic ice extents (Mysak and Manak, 1989), and in mean ice thickness in an adjacent portion of the Arctic Ocean (Wadhams, 1990) are alternatively interpretable in terms of either global warming or other unrelated processes (McLaren et al., 1990). Likewise, although a recent review of Canadian ice thickness monitoring data (Brown and Cote, 1990) showed no definitive thinning trends in nearshore fast ice it was concluded that the strong dependence of ice thickness on local and regional snow cover accumulations precluded any conclusions regarding evidence for global warming impacts.

It is clear, however, that the regional changes observed over at least the last three decades are at variance with the results of most GCM simulations which show (Figure 17) a warming of the study region during the winter in preference to other areas and other seasons along with decreasing winter ice extents. Instead, observations indicate that cooling has predominated in the region since the 1940's coincident with first an overall hemispheric/global cooling and then, since approximately 1970, simultaneously with a hemispheric/global warming. The predicted decreases in ice extent were definitively observed only during the anomalous global cooling of the 1960 decade. Ice extent increases were confined to the subsequent period of global warming.

In part, the differences between simulated and observed behaviour may be associated with the fact that most of the simulation results were derived as equilibrium responses to a simultaneous doubling of pre-industrial CO_2 levels. With present CO_2 concentrations only representing a total anthropogenic increase of slightly more than 20%, it is possible that the actual regional and global responses are, as yet, too small to be observed in the presence of other sources of variability.

Transient model simulations such as that carried out by Washington and Meehl (1989), do more closely approximate the evolution of the present atmospheric composition. Thus, it should be recognized that, as noted above, the mean temperature changes in years 26 to 30 of the Washington and Meehl (1989) simulation are negative over much of the Arctic, including all but the westernmost part of the study region. However, the features of this change, namely a weakening of the Icelandic Low and significant ice advance in the GIN Seas, most closely resemble observations during the anomalous 1960 period associated with overall global cooling. The transient results are almost directly opposite in sense to the changes observed during the last two decades of overall global warming when cooling and higher ice extents and warming and lower ice extents have been dominant, respectively, in our study region and in the GIN Seas. Also during this period, the Icelandic Low pressure centre has been, on average, strengthened.

At present, because of the relatively brief duration of the comparisons, it is not possible to decide whether the cited discrepancies between observations and simulated results represent fundamental errors in the latter or masking of the global warming effects by other time-varying phenomena. Certainly, in view of the significant differences noted (Washington and Meehl, 1989) between the $2XCO_2$ equilibrium and the transient simulation results, the rate and mechanisms of oceanic adjustment appear to be major factors in determining the nature of warming-induced changes in the critical Icelandic Low pressure feature. On the other hand, it would appear that at least as much significance should be attributed to the fundamental limitations of present GCMs. Particularly important in this respect is the limited or nonexistent capability of present models to adequately simulate the advective processes which, on the basis of the results discussed in Section 3.2, may be key determinants of ice extent variability. In view of the strength of the atmospheric/sea ice feedback mechanism, neglect of these processes undoubtedly introduces errors into not only the simulated ice extents, but as well into the atmospheric- (including wind-) variations also shown in Section 3 to be major influences in both the upstream and downstream components of ice variability off Labrador and on the Grand Banks. Possible impacts of advection on large scale global parameters such as temperature were also identified in a recent study (Ledley, 1991) of the dependence of ocean-atmosphere energy exchanges on the thinning of the polar ice cover by meridional ice transport.

Proper incorporation of advection requires the development of sub-models which successfully explain the origin of past ice cover changes. Inevitably, this development will require accounting for both the thermodynamic and advective contributions to the change process. Obvious points of interest are the oceanographic processes in the North Atlantic which are known (Dickson et al., 1988) to affect the temperature and salinity of the near surface waters throughout the sub-polar gyre of the North Atlantic Ocean. Past research into large scale ocean circulation, including global ocean circulation modelling, has focused on the understanding of the ocean dynamical response to large scale wind patterns. The reduction in general wind forcing associated with global warming as deduced through atmospheric GCM studies, would be expected to weaken the major ocean currents (eg. the North Atlantic Current and Gulf Stream in the N.W. Atlantic). However, the circulation features of most interest to this study are the buoyancy-driven shelf currents (eg. the West Greenland, Baffin and Labrador Currents), as discussed in Section 2.2), since it is these shelf currents which carry sea ice and icebergs southward along the east coast of Canada.

A better understanding of the ocean dynamics over the shelf regions, incorporating buoyancy fluxes resulting from ice melt and land runoff, is needed to properly understand the role of the ocean in responding to climatic changes. Recently, these research questions have begun to be addressed (Ikeda, 1990; Yao and Ikeda, 1990; Mysak et al., 1990; Ikeda, pers. comm.; Mysak, pers. comm.).

In the study region itself, some attention should at least initially be given to the role, if any, of side channel exports in determining the Davis Strait ice extents which appear to be representative of upstream influence on downstream ice behaviour. At present, there is little understanding of the extent to which these exports are determined by the regional wind fields or other larger scale Arctic factors. Their potential sensitivity to thinning of the mobile Arctic pack ice cover, which might be anticipated at some stage of the warming process, could be expected to increase channel ice mobility, which, in turn, to a certain point, could increase ice exports to Baffin Bay. This increase could produce ice extent increases not unlike those which appear to be associated with the present episode of overall global warming.

Detailed documentation of past and future variations in the upstream and downstream ice cover and in the accompanying pressure, wind and temperature fields would appear to be necessary to both assist in the improvement and refinement of GCMs and to facilitate implementing model extensions applicable off eastern Canada. Such contributions, along with the results of accompanying studies of the involved physical processes, should clarify the significance and underlying origins of the apparent post-1970 trend toward continued global warming coincident with unchanging, or very possibly, worsening ice conditions in the study region.

A reasonable objective for an effective assessment program would be the identification of the likely form and pace of a future progression from the present poorly understood regional response to global warming toward the simpler conditions which might be expected on the basis of virtually every equilibrium CO_2-doubling simulation. The latter conditions, consistent with simple expectations in terms of increased energy storage in global surface layers, correspond to definitively higher regional temperatures and almost universal ice retreat.

Effects of Atmospheric Warming on Fast Ice - A Case Study

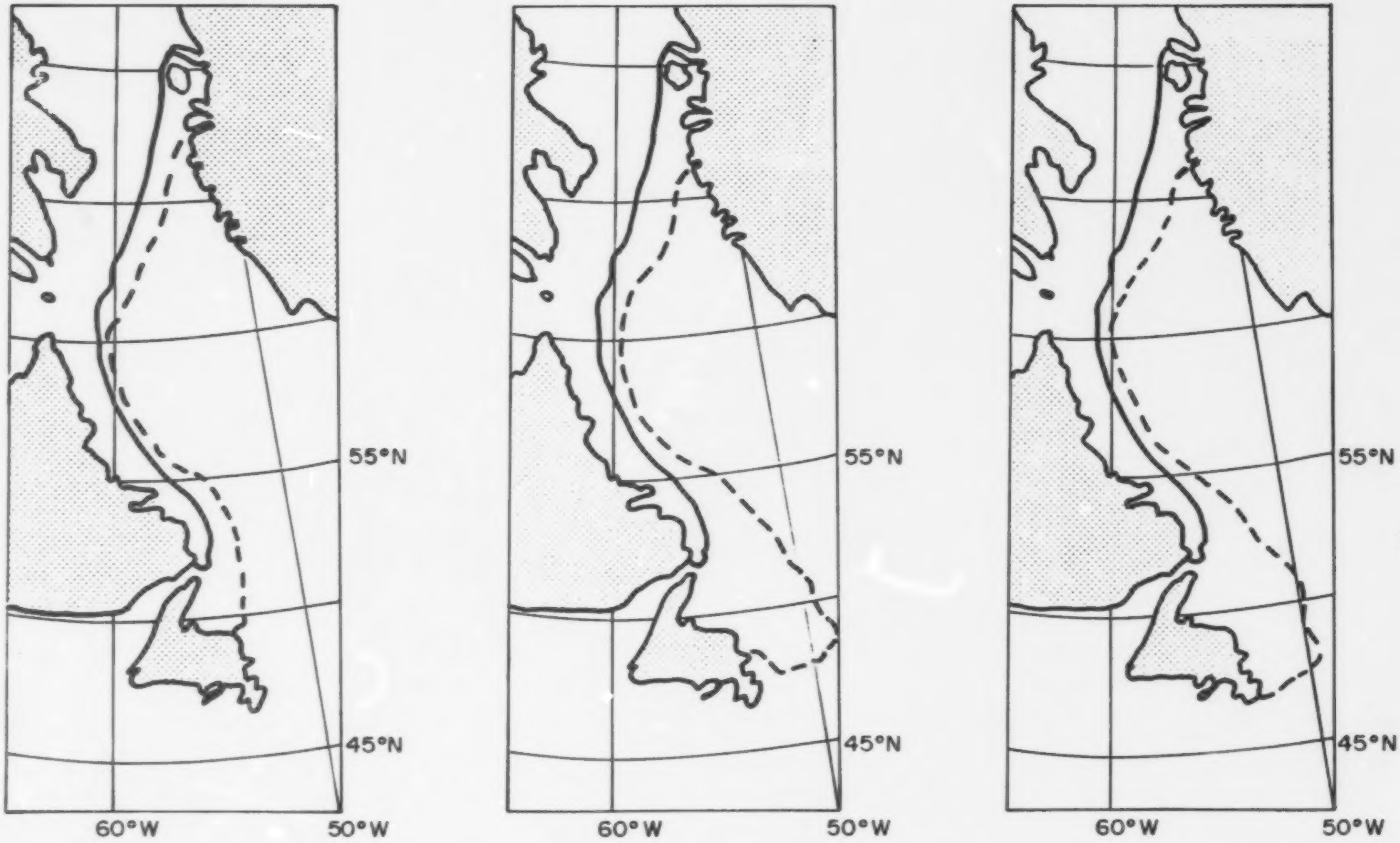
To obtain rough estimates of the rates likely to be associated with such a retreat and their relationships to regional warming, calculations were carried out for the thickness of sea ice which begins to form in early October at 74°N in western Baffin Bay, and moves south to southern Labrador by the following spring. Increases in ice thickness were estimated using the Bilello (1960) freezing-day relationships in a Lagrangian frame moving southward with the ice at the observed 15 km/day mean drift rate. A 10 cm snow cover was assumed and temperatures were set equal to the monthly means recorded at the closest coastal monitoring station over the period October-March. The estimated thickness of the ice off southern Labrador at the end of February was 125 cm, very approximately near the upper limit of undeformed first year ice thicknesses observed in March when the mean local temperatures (at Hopedale) are -2°C.

Subsequent use in the same manner of the Bilello (1960) melting degree day relationships together with mean April and May sea surface temperatures in the South Labrador Marine Grid (Danard, 1977) indicated that the thickness of this ice would be reduced to zero by May 25. As a consequence, and to facilitate estimation of warming-induced boundary shifts, the mean June 1 mean ice extent boundary as compiled by the U.S. Navy (NOC, 1986) was assumed to be approximately representative of the extents on the "disappearance date" of the tracked 74°N ice.

Year round elevations of 2°C, 4°C and 6°C were then assumed in the monthly temperatures along the track of the southerly drifting ice. These changes were found to reduce the end-of-February ice thicknesses to 105, 87 and 65 cm respectively. Subsequent use of the melting degree day relationships yielded disappearance dates for these thinner ice fields of April 10, March 18 and March 8 respectively.

Using our assumption that the present June 1 mean ice boundary is representative of mean ice extents on all future mean disappearance dates, the changes in mean ice extent under the three different levels of warming are indicated in Figures 28a-c. These figures show the corresponding present mean ice extents and the assumed universal "disappearance" boundary associated with the dates upon which, in each case, ice originating at 74°N in October may by assumed to have been reduced to zero thickness.

This approach ignores significant differences between the heat exchanges characteristic of nearshore fast ice (from which the Bilello (1960) data sets have been derived) as opposed to offshore pack ice. Nevertheless, in the absence of detailed treatment of oceanic heat fluxes and regional- and climate-dependent snow accumulations, etc., our admittedly crude estimates suggest that the effect of atmospheric warming in our areas of interest would be to shift the southern ice boundary northward by roughly 2.5°, 4.5° and 5° in latitude for the 2°, 4° and 6°C warming cases respectively. As well, the width of the spring pack at the southern end of Labrador would be reduced by approximate factors of 2, 3.5 and 3.5 in these same cases. Although the temperature data used in these calculations are not exactly appropriate for regions near Greenland it should be noted that similar northward shifts in the boundary would be anticipated in that region.



a. b. c.

Figure 28 Comparisons of present day and future mean ice extent boundaries corresponding to: (a) April 15, with a long-term uniform 2 C atmospheric warming; (b) March 15, with a long-term uniform 4 C uniform atmospheric warming; (c) March 1, with a long-term uniform 6 C uniform atmospheric warming. In each case the future mean boundary on the specified date is indicated by the solid line curve representing the universal "disappearance" boundary (section 4.2) while the present day mean extent on the same date is represented by broken line curve.

The estimated shifts for the 4° to 6°C temperature elevations are comparable to the present natural range of boundary position variability. In other words, the effect of global warming, in this very simple thermodynamic treatment, is approximately equivalent to a transition between the present maximum and minimum ice extent configurations in the spring period. Some evidence for this order of magnitude equality between the present natural climatic variability and the deterministic variations expected as a result of warming can be found in the modelling results of Herman and Johnson (1978). The latter results were obtained by running the GISS global climatic model with two sets of ice limits and inputs, representing the maximum and minimum winter ice extents over the period 1961 to 1977, but with other input parameters unchanged. It was found, for instance, that a minimal ice extent in Davis Strait corresponded to a mean surface temperature rise of 3-4.5°C, which is similar to the presently expected magnitude of greenhouse warming. Although it was possible to obtain a temperature increase by forcing a climate model with reduced ice extent, it is not necessarily true that forcing the same model with the obtained temperature rise would produce a true original equilibrium ice extent input. Existing models are not reversible in this sense. Nevertheless, this result gives some support to the idea that the expected future retreat will be of roughly the same magnitude as the amplitude of natural fluctuations over the last 20 years or so.

Modification of these results can be expected to occur with anticipated increased regional precipitation rates which, through the increased insulation provided by thicker snow, should further thin the ice cover and hence most likely decrease spring ice extents. Likewise the anticipated increase in the frequency of southerly winds, associated with a weakening or shift (e.g. westward migration) of the Iceland low, would both retard the southward drift of upstream ice and push the warm Atlantic water further north. In both cases, this would tend to further reduce regional ice extents. On the other hand, a freshening of the shelf surface waters resulting from ice melt and increased land runoff could have mixed effects, accelerating buoyancy-driven southward-flowing currents and increasing water column stability and thus allowing more near-surface warming. Likewise, slightly smaller amounts of retreat might be expected if the summer warming is significantly smaller than that which occurs in winter as indicated by the ensemble results (Wigley et al., 1990).

The details of the progression between the present still poorly defined regional trend and something resembling the above-sketched longer term changes will require broad advances in the understanding of the North Atlantic and regional atmospheric and oceanic circulation with particular emphasis on interrelationships with the sea ice cover.

4.3 Climate Change and Iceberg Severity

Relationship of Iceberg Severity to Sea Ice Severity

On the basis of the discussions of Sections 2.4 and 3.3, separate contributions to the observed interannual variability in seasonal Labrador/Grand Banks iceberg severity have been associated with the processes which govern berg production (calving) at the principal glacial sources of Greenland and in the en route attrition of berg numbers and masses. On the basis of the observed correspondences, and process understandings, major portions of past variabilities in iceberg numbers have been explicable in terms of sea ice extent. Specifically, the attrition of bergs in the waters off southern Labrador and on the Grand Banks is closely related to quantity of sea ice along the iceberg transects. With few exceptions, year to year variations in the upstream berg supply associated with corresponding changes in calving rates or rates of attrition north of 55°N appear to contribute downstream differences which are smaller than the individual ranges of the approximate severity categories definable within the limitations of past annual counting accuracies.

Climate-induced changes may thus be expected to follow closely those which would occur in the accompanying south of 55°N spring ice extents approximately, as indicated by the empirical comparisons of Figure 15.

Possibly significant differences could however be expected if large changes were to occur in either or both the upstream berg attrition rates or in the rates of berg production or calving.

In the first instance, the most relevant changes would be alteration of currents, winds and temperatures which would affect seasonal water temperatures, ice extents and sea states which, in turn, govern berg dissipation, grounding and reflotation rates. Although the details of the net effect of these changes are undoubtedly complex, a reasonable guess would be that the net impact of change on upstream attrition will follow that observed in downstream attrition. In other words, reduction in downstream spring ice extents will probably coincide with overall lower upstream extents during the non-winter months when coastal shelf entrapment and deterioration affect southerly berg fluxes. Short of a massive regional cooling response to further global warming which would increase the longevity of the coastal fast ice shelves and hence decrease entrapment and attrition, it is hard to visualize mechanisms for upstream changes which will not further reduce downstream berg severities.

As a consequence, the potential for an alternative increase in berg severity lies largely in the possibility that an overall global warming could produce a dramatic increase in berg calving rates. Such increases would have to be either sufficient to more than compensate for the increased attrition rates which might accompany any amelioration of regional sea ice severities (Wadhams and Dowdeswell, 1990) or, alternatively, would have to occur prior to the start of such a process.

Certainly, in the case of the long term universal warming considered at the end of the previous section, the approximate associated reductions in spring ice extents can be expected to be comparable to those presently associated with alternative high and low values of the January Davis Strait ice index. As a result, in view of the observed correspondences (Figure 15) such reductions might be expected to produce comparable changes in berg severity. This suggests that future high berg severity levels may resemble those presently identified with low severities, i.e. less than 200 bergs, south of 48°N.

Iceberg Production and Climate Warming

As a consequence, even though sea ice extent variations remain the dominant influence in determining berg severities, quantitative assessment of future berg numbers south of 48°N still requires estimation of the magnitudes and time scales which could be associated with climate-related changes in berg calving and production rates.

Calculation of the latter changes were carried out during the present study (Wadhams and Dowdeswell, 1990) (included in Appendix B) using the assumption of equilibrium between a representative West Greenland glacier, the Jakobshavns Isbrae (at 69°N) and the atmosphere. These calculations were based upon the balance flux approach previously used by Bindschadler (1984) to estimate possible future glacier melt contributions to global sea levels. In this case, however, a 1300 m present-day equilibrium altitude (the altitude of zero net annual glacier ablation/accretion) was assumed while retaining the same equilibrium plastic flow relationship between surface height and distance along the glacier. The balance or ice flux at the lower end of the ice drainage basin was set equal to the rate of iceberg production. The range computed for the present-day rate (encompassing the uncertainties in the estimated size of the Jakobshavns Basin) (Figure 29) is in good agreement with both observed calving rates and a previous estimate by Bindschadler (1984), as well as with a more recent rate estimation of 42 km^3/yr (Echelmeyer, pers. comm.). These results suggest that the Jakobshavns Isbrae is responsible for 15 to 25% of total West Greenland berg production.

On the basis of this agreement, the balance flux method was used to estimate the likely direction and magnitude of the calving rate changes to be anticipated, in equilibrium, in a warmer world. Two specific warming scenarios were tested corresponding to a 10% rise in annual precipitation and alternative 500 m and 1000 m elevations of the equilibrium level altitude. On the basis of the varying reported values for the gradient of surface temperatures with respect to altitude at 70° to 71°N (Ambach, 1982) and at 76.5°N (Benson, 1962), these equilibrium level altitude changes are very approximately equivalent to 3°C and 6°C regional temperature rises, respectively. Such changes are in accord with the 4°C and 6°C temperature elevations predicted by Hansen et al. (1984) for southern and northern Greenland, respectively, under a doubling of pre-industrial concentrations of atmospheric CO_2.

The resulting ranges of calculated calving rates differ significantly from present day estimates only for the most extreme of the two assumed scenarios (i.e. corresponding to a 6°C temperature rise). It is important to note that in this extreme warming case, the anticipated changes correspond to roughly a 50% reduction from present berg production rates. Thus, in equilibrium, a decrease in basin size arising from rises in the equilibrium level altitude more than compensates in terms of calving rates for the increased accretion levels expected at higher elevations as a consequence of both the greater moisture-carrying capacity of warmer air and decreased sea ice extents.

It should be noted, however, that the utilized calculation method does not account for lag times between changes in precipitation and ablation rates and the propagation of effects downstream in the glacier field. At the velocities of less than 50-100 m/yr observed in the upper part of the Jakobshavns basin by Malzer (1964), it will take several thousand years for ice build-up in the upper basin to be reflected in changes at the coast (cf. Radok et al., 1982) assuming kinematic wave behaviour does not take place (cf. Nye, 1960, 1963).

The time scale on which the calculated changes would operate is thus uncertain due to such lags between the ice sheet response and changes in the mass inputs. Nevertheless, the direction of change in iceberg production at equilibrium appears to be toward a decrease in iceberg calving for the more major climate change specified. Little change is anticipated from the second of the tested scenarios corresponding to a smaller (500 m) equilibrium level altitude elevation (approximately equivalent to a 3°C year-round warming) and the same 10% precipitation increase. Nevertheless, it is important to note that we cannot, at present, rule out the possibility that transient ice flow effects and local outlet glacier responses, neglected in the equilibrium treatment, could lead to short-lived increases in iceberg production.

Past attempts (Bindschadler, 1985) to estimate transient increases in calving rates using observations made on a single grounded Alaskan glacier (Sikonia, 1982; Brown et al., 1982a) did produce large increases ranging from roughly 80% to 200% of present day rates. Such

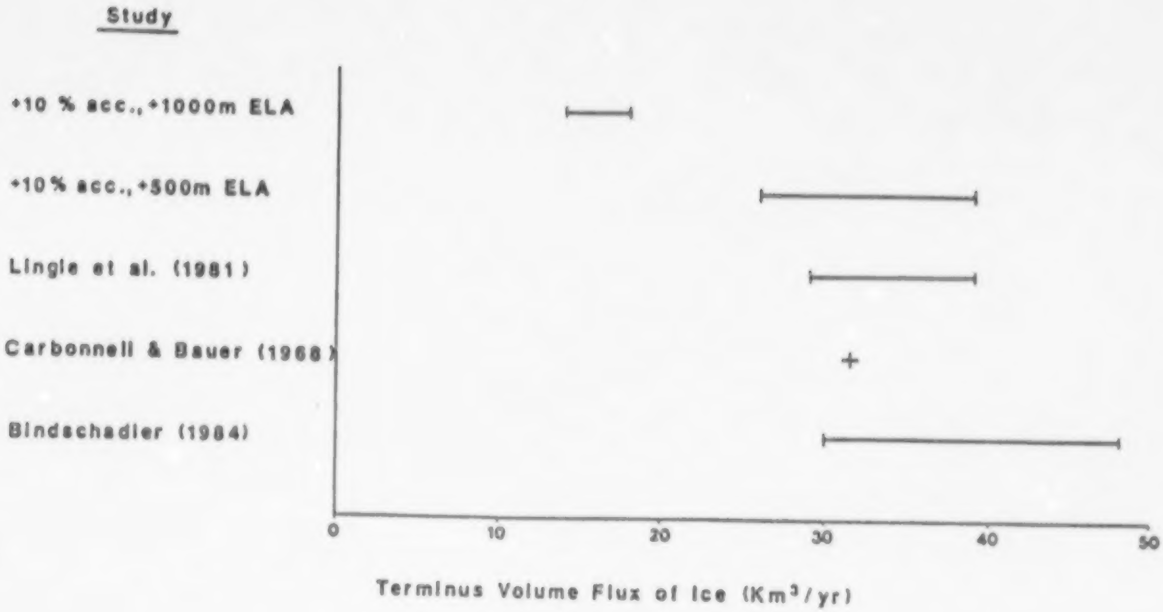


Figure 29 Calculated and observed values of the volume flux at the terminus of Jakobshavns Isbrae. The upper three values were calculated in the present study for, respectively, the present day conditions and the altered conditions described in the text (Section 4.3).

increases would appear to be associated with alteration of the resistance to flow at the substrate interface. However, because of fundamental differences in glacial type, these extrapolations are not clearly applicable to conditions at the floating lower ends of West Greenland iceberg-producing glaciers. As well, recent results (Echelmeyer and Harrison, 1990) obtained from studies of the Jakobshavns Isbrae indicate the absence of significant seasonal dependence in the speed of the surface glacial ice advance. Since the amount of meltwater available for basal "lubrication" is a strong function of season, these observations call into question the likelihood of meltwater-induced transient increases in calving rates. Yet, it must be recognized that it may be possible for the Greenland glaciers to respond to warming in the form of a series of surges and slow-downs which occur on timescales much shorter than those associated with our calculated equilibrium response. Unfortunately, however, present knowledge of transient behaviour is such that these potentially important refinements remain as primary topics for further investigation and research.

Some work on dynamical- as opposed to quasi-equilibrium-glacier responses has been carried out by Nye (1960, 1963) who showed that, if a change in accumulation rate is imposed on all or part of a glacier, the effect is initially propagated downstream in the form of a kinematic wave. This wave would pile ice up against the lower end of the glacier and result in a transient increase in thickness and/or a surge forward of the glacier terminus. Clearly this would lead to a transient increase in iceberg calving rates. Time scales for kinematic wave propagation are in the range of 3 to 30 years, depending on glacier size. Yet, Paterson (1981) has pointed out that glaciers are constantly adjusting to local changes and variations in accumulation rate and temperature, and that there are few clear recorded instances of kinematic waves. The result of such a wave, relative to the equilibrium adjustment predicted by the flowline model, would be an initial excess, followed by a dearth of icebergs when the additional mass has been discharged. We feel that the time scale of the climatic change considered places it just beyond the point at which it can be considered a step function impact, capable of generating a kinematic wave. This transient response would, therefore, not appear to play a major role in determining future calving rates.

Nevertheless, further consideration of the transient response problem appears to be essential in view of the above cited uncertainties in the time-scales of future regional climate changes. Thus, for example, if the postulated 3°C to 6°C rises in the regional summer temperature do occur prior to a change from the most recently observed regional fall/winter warming response pattern (i.e. colder air temperatures and unchanged or higher ice extents) it is possible that transient surges in berg production could lead to corresponding percentage increases in downstream berg numbers. On the other hand, in view of the apparent extreme sensitivity of attrition rates to sea ice extents, such transient surges in calving rates would be unlikely to have significant downstream effects if they follow or are coincident with the projected eventual start of the northward retreat of the mean annual regional ice edges.

There is some evidence (Zwally et al., 1989) that the ice sheet of southern Greenland has been thickening at a rate of roughly .23 m/yr since the mid 1970's. This result is consistent with expectations of increased precipitation associated with the warmer summer temperatures observed during this period. Although, to the best of our knowledge, no accompanying dramatic surges in calving rates have been reported, clearly future monitoring of berg production is required to establish the timescales and magnitudes which could be associated with short-term production increases.

5. SUMMARY AND RECOMMENDATIONS

5.1 Summary of Key Features of Regional Ice Environment

It is evident that significant progress has been made, in recent years, toward understanding and quantifying the processes which most immediately determine sea ice and iceberg severities in the areas of interest. The central role of the spring Labrador ice in protecting its contained berg population and, in effect, determining probabilities for berg survival, is consistent with both present understandings of berg deterioration and the significant cross-correlation of annual ice extents and berg severities.

Likewise, the importance of the January Davis Strait ice extent in determining spring ice extent off Labrador and Newfoundland, as modulated by March/April wind variability, is understandable in terms of the southerly advection of the regional ice and accompanying movements and heat transport changes in the underlying water masses.

The interannual variations in the Davis Strait ice extent intermediary also appear to be closely related to the strength of northerly or northwesterly wind forcing in the northern portion of the study region during the early and mid-winter months and hence to variations in the associated atmospheric pressure distributions. It is possible, but less certain (from anomalies in the observational record and physical considerations), that contributions to interannual variability may also be associated with variations in sea ice and water exports from regional side channels connecting the study region to the greater Arctic Ocean.

Although upstream berg data were much more limited, accumulated results strongly suggest that the major source of attrition for the huge population of annually produced bergs is associated with grounding and entrapment on the shallow continental shelves of the region. Seasonal fast ice is an important factor in this process which, in effect, acts to supply a relatively distinctive pulse of bergs to Davis Strait over roughly the November-January months. The resulting end-of- January Davis Strait berg population eventually moves into the areas of interest during the April-June period usually associated with the iceberg season. Present evidence is consistent with the likelihood that variations in berg production and upstream transport/attrition are of lesser significance relative to downstream ice-controlled attrition processes.

It is quite clear from the modern data records that high sea ice extents in Davis Strait in January and south of 55°N in April are both necessary precursors of subsequent high, south of 48°N iceberg numbers. The demonstrated linkage between the two sea ice extent parameters and their relationships in time and space suggest that an important portion of the physical connection between the global climate and southern ice and berg severities must be sought further upstream at times prior to the spring-summer season of traditional hazard.

Reviews of the reasonable to excellent quality ice extent data available for the past three decades and of much poorer quality indirect data, dating back to, at least 1920, reveal irregular patterns of interannual variability including some quasi-periodic variations over approximately decadal time scales. However, the most pronounced feature observed in sea ice severity is a very marked decrease in ice extent in the 1960's, and possibly, to a lesser degree, in the 1950's. The 1960 decade reduction was followed by two decades of above average ice extents.

Over longer time scales of decades or more, it is clear that the variations in sea ice extent appear to be coherent throughout the larger study region (Figure 1). Moreover, the interannual variability in sea ice is well correlated with related meterological (air temperatures and northerly wind components) and oceanographic (heat content of the continental shelf area water column) parameters. This consistency within the study area implies that the observed longer term variability is related through large scale linkages to influences located outside of the region. Previous research efforts have identified several possibilities, including large scale atmospheric systems and, in particular, the Icelandic Low centred between Greenland and Iceland which is involved in manifestations of the North Atlantic Oscillation (NAO). Interannual variations associated with the NAO are related through atmospheric teleconnections to other large scale atmospheric-ocean patterns, such as the El Nino-Southern Oscillation (ENSO). Another possible large scale external linkage which may affect the study area is the subpolar gyre within the North Atlantic Ocean, which represents a basin wide cyclonic recirculation connecting the Greenland-Iceland-Norwegian (GIN) Sea with the Labrador Sea. The upstream GIN Sea is a particularly important mixing region on a global oceanic scale, since it is the site for both major exchanges between the Arctic Ocean and the rest of the world ocean, and for the deep ventilation of heat from the abyssal depths.

Either on or beyond the extended regional spatial scales of the North Atlantic - Arctic region, there also may be some important linkages to global scales. The sharp reduction in ice extent in the 1960's, which occurred in all parts of the study area, coincided with a period of global and hemispheric cooling. For the two decades since the 1960's, sea ice extent has been at generally high levels, accompanied by low winter temperatures and strong northerly winds. This behaviour has coincided with a resumption of overall global warming. The opposite signs of these trends in winter conditions off the Canadian East Coast and in larger scale (global, hemispheric and Arctic basin) mean temperatures are presently not consistent with expectations as deduced from equilibrium $2XCO_2$ GCM simulations or from related transient simulation which more realistically duplicate atmospheric modification and environmental adjustments.

Overall, even with the present day, very incomplete understandings of regional, extended regional and global processes, there is no evidence that the accumulated effects of anthropogenic atmospheric modification have contributed to amelioration of ice and iceberg conditions off eastern Canada. On the contrary, since roughly 1960, winter regional temperatures and ice severities have been directly out-of-phase with corresponding global trends. In the absence of information relevant to a change in a presumed linkage between regional ice severity and global temperatures, one would have to assume that, in the immediate future, ice and iceberg severities will either hold at their presently high levels as observed over the past decade, or perhaps even worsen with continuation of the present global warming. Of course, oscillations on roughly decadal time scales will continue to occur due to natural internal atmospheric ocean phenomena, such as variations in the ENSO and North Atlantic Oscillation Index. Moreover, as in the past (Hansen and Lacis, 1990), temporary perturbations of the atmosphere by volcanic activity and other climatic factors might be expected to produce intervals of milder regional conditions. Consistent interpretation of the variability in global air temperatures will require the inclusion of such shorter-term changes in atmospheric forcing, as well as some examination of the origin and stability of the presently observed out-of-phase relationship between the regional and global responses to atmospheric change.

Nevertheless, it is possible that, in the long term, high levels of greenhouse gas-induced global warming corresponding to 2° to 4°C rises in the annual global mean temperature will

produce an opposite response corresponding to warming and ice retreat. This more intuitively agreeable situation of less sea ice in a warmer world would reduce the numbers of icebergs seen in southern regional waters, due to increased attrition of the bergs in the absence of the protective effects of sea ice. It would then be expected that the numbers of icebergs would be reduced to, at most, levels currently associated with low severities.

An alternative increase in iceberg severity in the long term case would appear to require the presence of a mechanism whereby calving rates could increase fast enough to compensate for the increased attrition associated with lower sea ice extents. In the absence of any present evidence for such increases either in the equilibrium calculations of Wadhams and Dowdeswell (1990) (Section 4.3), or from observed dependencies on meltwater generation (Echelmeyer and Harrison, 1990), it is unlikely that future regional conditions will include simultaneous low seasonal ice extents and high iceberg severities.

A description of the transition between the immediate future scenario (i.e. continuing relatively high; out-of-phase [with global temperature] severities) and the even more speculative much warmer world or other scenarios poses the major challenge for climate change assessment. The transition period would occupy the greater portion of the next century, and, quite conceivably, could be characterized by a number of discontinuities in behaviour or "mode-switching" in the atmospheric, oceanic and ice components of the environment. Anticipation of these changes is likely to be achieved only by considerable advances in present understandings of both global and regional processes.

5.2 Outline of an Assessment Approach

On the basis of the foregoing summary, it would appear that meaningful anticipation of future regional climates must, at present, be confined to periods in the next few decades when we can assumed with some justification that the recently observed out-of-phase regional/global climate relationships will continue. In this period, we might expect either little change or perhaps worsening relative to the colder, icier regional conditions characteristic of the past two decades. An opposite extreme of regional behaviour consisting of the drastic elevations of temperature predicted by GCMs for $2XCO_2$ equilibrium conditions would produce the kind of universal sea ice retreat which could reduce maximum future ice and iceberg severities to levels comparable to, or lower than, historical low values.

The primary task of future regional climate change assessments must be to obtain reliable, realistic estimates of the rate and characteristics of changes from the present regional response to continuing warming on larger (global) scales. At some point, estimates will be required for the relative timing and occurrence probabilities for the second specified scenario (regional warming and ice retreat) along with identification and evaluation of other, more refined scenario alternatives.

Success in these regards will require significant improvements in understandings of the processes which control sea-ice extent through linkages with atmospheric and oceanic factors.

Studies which rely solely on statistically inferred relationships between sea ice and other parameters are not likely to be very helpful, given the short duration of historical ice data relative to the time scales associated with both natural interannual variability and anthropogenically-induced changes. Prospects for the use of proxy indicators, which might extend the data set over much longer periods, do not appear promising for sea ice within the study area.

The first step towards this general objective would appear to be to more carefully analyze the historical data of the past three decades or more. These analyses should focus on the key ice, atmospheric and oceanic parameters identified in this study, and be directed at deriving a better understanding of physical mechanisms affecting regional and subregional ice extent. Although the initial parameters of interest, for our purposes, would be the winter and spring ice extents in Davis Strait and off Labrador, respectively, much effort would have to be given to identifying, if relevant, anticipated changes in ice conditions further upstream, and, in particular, in the regional wind and air pressure climatology and in the stability of the North Atlantic oceanographic circulation and mixing processes.

Ideally this program should lead to an understanding of ice extent variability (probably in the form of a regional ice model) which is self consistent with corresponding larger-scale atmospheric and oceanic parameter distributions as predicted by improved GCM simulations. As a guide to model development, and in recognition of the absence of equilibrium among the interacting systems, considerable effort would have to be given to updated empirical analyses which would seek evidence for significant changes in regional response patterns.

Building upon the results of these studies, the derived-physical linkages would be used to assess as much as possible, the degree to which the past regional responses, as observed at least over the past three decades, represent a characteristic response to change in overall atmospheric composition. This determination is of most immediate interest to better defining the extent of tangible present-day effects of the global warming process. Doubts of the reality of such effects have, to some extent, focused on the slightly smaller than predicted (by $2XCO_2$ equilibrium models) increase in mean annual global temperature, as well as on the failure of the warming to be preferentially concentrated in the polar regions, again as predicted by the models.

Undoubtedly, the task of unambiguously identifying the source of regional climate change with global warming related- as opposed to other-mechanisms is not an easy one. Nevertheless, the efforts necessary for such an identification, namely a detailed examination of the possible, at least, intermediary sources of past regional behaviour, are such as to produce the kinds of knowledge which would assist the development of both GCMs, which must provide the correct future larger scale forcing, and of regional models, which would translate such forcing into meaningful climate assessments in our areas of interest.

Progress in such a program will obviously require close coordination of regional studies with external research directed at advancing knowledge of climate change responses on both the global scale and in adjoining regions such as the GIN Seas. Recommendations are given below (Section 5.3) for the more immediate research requirements specific to the regional study component directly relevant to East Coast ice climate change.

5.3 Recommendations

Overall Approach

The most immediate requirement for progress toward an assessment of future ice severity changes is the identification and development of understandings of the mechanism(s) which controls present and past variability in the key January Davis Strait ice extent parameter, and subsequently in the closely linked spring ice extents south of 55°N. This understanding should include a capability to anticipate modification of present behaviours by continuing atmospheric change and should, as well, allow development of self-consistent modelling of future atmospheric parameter (i.e. temperature, pressure, winds) distributions.

Lesser, but still significant, priority should be given to filling important gaps in present knowledge of iceberg production, transport and deterioration processes. A particularly critical requirement is for quantitative estimation of timescales and magnitudes associated with transient or non-equilibrium changes in iceberg calving rates under the anticipated pace of atmospheric warming.

Sea Ice Interannual Variability

Present efforts (Ikeda et al., 1988; Ikeda, 1990; Peterson and Prinsenberg, 1990) to understand sea ice variability in southern areas (south of 55°N) should be expanded to include the ice covers of the upstream regions which appear to be, directly or indirectly, principal determinants of downstream severities. The needed information for establishing the quantitative importance of atmospheric- and, possibly other- influences (i.e. upper ocean variability, side channel ice dynamics) is contained in presently available historical data collections. Key components of the relevant databases are: an 18 year duration archive of NOAA AVHRR imagery held at the Environment Canada, Atmospheric Environment Service Ice Centre in Ottawa and the historical set of areal ice extents available from ship-based, aerial and, later, satellite-measurement platforms extending back to the 1950's (eg. Walsh, 1978). Automatic processing software is now available for extracting ice velocity fields and other geometric- and extent-data from selected satellite image sequences. The comparisons of such data products with contemporary atmospheric temperature- and geostrophic wind-fields, and parameterization for use with earlier pre-satellite derived ice extent records, should provide a basis for improved process understandings. Such understandings of control mechanisms could guide the refinement of ice models to more accurately account for advection in the determination of ice extents and volumes.

As well, further research into ice variability in the downstream regions is required. One important knowledge gap for the purposes of climatic impact studies is the role of cross-shelf winds in changing the ice extent. Further understanding is particularly relevant in the case of the easterly winds during late winter or early spring which appear to modulate the sea ice extent off Newfoundland and S. Labrador. Knowledge of the mechanisms and controlling parameters, which lead to abrupt, and usually irreversible, late season reductions in sea ice extent is important in understanding the interannual variability of the sea ice, and iceberg severity, in southern regions.

In the historical analysis of ice data, initial focus should be given to looking at extreme (high or low ice extents) years to facilitate the identification of important characteristics of individual annual behaviour, such as might be expected to be derived from: the fields of ice motion; positions of fast ice boundaries and thin ice/open water; frequency of cloud cover and relationships to geostrophic wind forcing and upstream water properties.

In the last regard, because of continuing uncertainties in the appropriate cause-effect relationships between water properties and sea ice extent in Davis Strait and off Labrador, all available data should be examined with regard to upstream water properties in regions as distant as the Greenland Sea and the side channels of Baffin Bay. Although most of this work should initially be based upon historical data, close contact should be maintained with future field programs in these areas such as the upcoming NSERC sponsored research in the North Water polynya region at the northern end of Baffin Bay.

Research programs directed at examining connections between ice extent and upper ocean modification (Ikeda, 1990; Yao and Ikeda, 1990; Mysak et al., 1990) are essential in improving our understanding of important ice-ocean processes. Particular emphasis is required on (1) improved understanding of coastal circulation dynamics associated with buoyancy driven currents driven by ice melt and land runoff, and (2) the deep water exchanges and ocean ventilation in the Greenland and Labrador Seas. Importance should be attached to conducting these research efforts using both modelling and improved data acquisition techniques. At present, ocean and ice-ocean modelling efforts are often limited by the paucity of input or verification data sets available for the ice covered waters off Eastern Canada.

Major deficiencies in existing data include vertical heat fluxes (both from depth within the ocean and through the air-sea or air-ice-sea boundaries) and ice thickness measurements. For heat fluxes, some first order estimates of long term trends in mean atmospheric heat fluxes could be derived from the use of historical bottle cast/CTD data in conjunction with a simple box model. Further progress may be made through the use of instrumented platforms operated from the sea-ice and a new generation of satellites planned for launch which should provide regional measurements of winds over the ocean.

Ice thickness data are required for two reasons. (1) as a possible means of monitoring and anticipating changes from the past- and present-day ice cover response to warming and (2) as a requirement for better modelling of ice dynamics, especially in the southern margins of the ice zone where the ice thickness is both highly variable and a key factor in determining the dynamical behaviour of the ice pack. Reliable, practical airborne measurement techniques need to be demonstrated and employed at regular intervals along fixed survey lines.

Assembly and further analysis of historical oceanographic data over longer time periods (1920's or earlier) would be helpful. A large amount of oceanographic data has been collected by European countries, especially Denmark and Germany, in the Davis Strait and Baffin Bay areas since the early part of this century. Recently, portions of these data sets have been assembled and analyzed through NAFO, the North Atlantic Fisheries Organization (eg. Myers et al., 1988). Further research of this type could contribute to better understandings of the interannual variability of the ocean and of possible linkages between the ocean and sea ice regime. Additional refinements of the studied ocean parameters themselves could also be beneficial. The Station 27 ocean temperature data presents a good example in this respect. Integration of anomalies over the water column should increase the robustness of the derived proxy indicator while still retaining the apparent and physically reasonable connection between earlier upstream sea ice extent and downstream water temperatures.

As well, the database underlying the Newfoundland spring ice index Miles (1974) also deserves further review. The recent work of Hill and Jones (1990) was an important contribution. Efforts at extending the historical record back further, and improving its accuracy, would be valuable.

Finally, considerable priority should be given to ensuring that the long-term historical time series of useful ice, oceanographic and atmospheric parameters are continued and extended where possible. For example, sea ice extent data collected from satellite and airborne sensors are essential for continuing the climatic change studies. Similarly, continuation of monitoring oceanographic conditions at Station 27 and along other standard NAFO transects, and ongoing air temperature measurements at coastal weather stations, represent a valuable climatic research resource. Care must also be taken to ensure that any technological changes are fully taken into account in order to provide measurements consistent with the long term historical data sets.

Iceberg Interannual Variability

Major uncertainties exist concerning the relative importance and timescales of transient berg calving responses and in the relative magnitudes and interannual variability of the three identified berg transport components.

In particular, present understanding of berg deterioration mechanisms and their respective dependence on ocean, atmospheric and ice parameters is very limited. More direct observations of iceberg deterioration are critically needed, including specific efforts to identify suitable stable, easily recognizable bergs for berg mass loss rate studies. Periodic visits by aircraft or ship and careful measurement of above-water dimensions and surface water parameters should provide, at least, lower bounds for realistic loss rate estimates. In addition to the acquisition of more and better data, improved theoretical and modelling efforts are needed to provide the knowledge required for significantly improving our understanding of interannual variability in the south of 48°N iceberg severity index.

As well, progress in understanding berg calving variability and atmospheric feedbacks would be assisted by documentation of the present and past relationships between sea ice extent, and precipitation and accumulation rates on the Greenland Ice Sheet. Most of these studies would require integration of existing or planned programs for monitoring ice sheet thickness (by radar altimeter) (Zwally et. al, 1989) and satellite monitoring (with Landsat and future all weather radar imagery) of glacier front plan views and iceberg numbers with continued, more detailed, data gathering on berg distributions including accumulations of seasonal survey data and the use of satellite-tracked beacons.

Given the major uncertainties in our present understanding of iceberg production rates, continued and expanded operational monitoring of iceberg calving from the major source glaciers would be useful. The radar-equipped satellites being launched in the first half of the 1990's make this type of data collection more feasible than previously.

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APPENDIX A

APPENDIX A

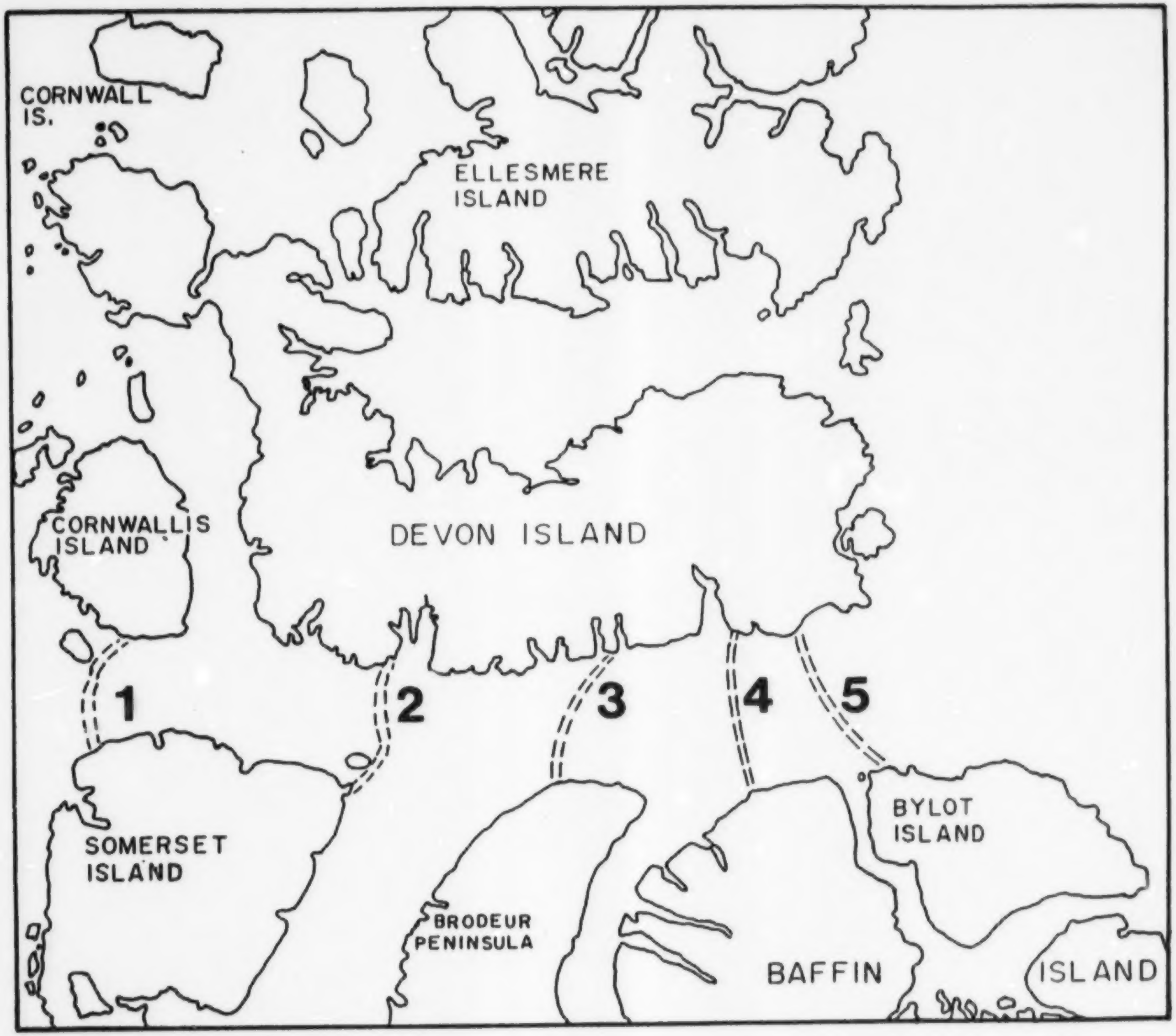
Winter Ice Export from Lancaster Sound

Observation of the winter ice cover in Lancaster Sound (Marko, 1981) indicates that ice edge stabilization tends to occur during the winter at one or another of several approximate locations along the channel length (Figure A-1). Prior to a final stabilization at a single permanent ice edge, usually in the January-March period, the stable edge shifts about among these positions as stable ice becomes mobilized and/or mobile ice ceases to move.

Once an ice edge is established, the fast ice to its west continues to thicken in approximate accord with the Bilello (1960) degree-day growth relationships, while the eastern mobile ice continues to drift eastward with typical velocities of 7 to 10 km/day. The drift is approximately uniform across the width of the channel (Marko, 1981), presumably, since the large-scale mechanical strength of the ice cover is large enough to preclude localized response to the spatial varying current patterns characteristic of the region (Fissel et al., 1982). The resulting drift tends to produce, because of continuing growth during drift, a wedge-shaped (in longitudinal cross-section) mobile cover with its thickest end immediately adjacent to Baffin Bay. Open water is usually present at the thin end of the mobile ice at the fast ice edge, except perhaps when ice is in its easternmost positions #4 and #5 (Figure A-1) when thick Baffin Bay ice may be brought up to the fast ice edge by the relatively shallow (20 km) intrusive loop of the Baffin current into the mouth of the Sound.

Estimates of the differences in exported ice volume were made assuming negligible exports for a stable November-January edge at positions #4 or #5 as opposed to a stable edge over the same period at the westernmost position #1. Uniform drift of 10 km/day was assumed across the approximately 100 km mean width of the channel and from the Bilello growth relationships and observations (Marko, 1981) the exported ice was assumed to be 1 m in thickness from November 15 through the end of January. This simple model suggests an additional 75 km^3 of ice could be expected to be exported into Baffin Bay prior to February if the fast ice edge were located in extreme western, as opposed to extreme eastern, positions.

It is apparent that the influence of Lancaster Sound ice exports on the end of January ice extent in Davis Strait is not dependent upon the exported ice reaching the latter region prior to February. The only requirement in this respect is that the accommodation of the exported ice into the Baffin Bay ice cover produces significant acceleration in the advance of the southern pack ice boundary. Clearly, accommodating the exported ice will produce rafting, ridging and hence mechanical thickening of the western Bay ice. Nevertheless, the internal ice stresses generated in the incorporation process should contribute to the acceleration of the southward ice flow in a manner similar to that expected in an incompressible ice cover.


CORNWALL IS.
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SOMERSET ISLAND
BRODEUR PENINSULA
BAFFIN
BYLOT ISLAND
ISLAND

Figure A1 Approximate positions of fast ice edges in Lancaster Sound and Barrow Strait (Marko, 1981).

APPENDIX B

CHANGES IN ICEBERG PRODUCTION FROM THE GREENLAND ICE SHEET IN RESPONSE TO GREENHOUSE WARMING

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January, 1990

1. INTRODUCTION

The scope of the present study is to investigate the problem of the impact of greenhouse warming on the production of icebergs from the Greenland Ice Sheet, and to estimate the magnitude of the change that will be produced by a specified climatic warming. This forms part of a larger project in which the modified production rate is combined with modified decay and drift rates to yield an estimate of the change in iceberg occurrence within the coastal waters of Labrador and Newfoundland.

Our underlying assumption is that over a period of a few decades the climatic warming will have an effect upon the Greenland ice sheet which lies within the limits of two scenarios:-

Scenario A : 500 m rise in equilibrium line altitude (ELA);
10% rise in precipitation.
Scenario B: 1000 m rise in equilibrium line altitude (ELA);
10% rise in precipitation.

In addition, local changes in precipitation in Greenland coastal regions are assumed to occur in response to sea ice retreat in Baffin Bay and the Greenland Sea.

The two scenarios are based on Bindschadler's (1985) interpretation of the predictions of the numerical model of Hansen et al. (1984), and correspond to the supposed effect of a doubling in atmospheric CO_2 concentration relative to 1900 levels. This is expected to occur some time after 2030. It should be noted that when the predictions of five current global climate models are compared (House of Lords, 1989) the estimate of a 10% rise in precipitation is common to all of the models but is a global average; the rise is proportionately greater at high latitudes and along mid-latitude depression tracks. We have included a 10% rise in precipitation in Scenarios A and B.

2. NATURE OF ASSUMPTIONS

2.1. Sea Ice Retreat and Precipitation

The first problem is how we should modify the distribution of precipitation to correspond to a modified pattern of sea ice extent which may occur under greenhouse warming.

A map of present-day accumulation over Greenland (Fig. 1, after Reeh, 1985) shows high values in the south, up the west coast as far as about 69°N, and up the east coast as far as about 71°N. There is another pocket of high precipitation in the coastal region north of Melville Bay, at 76-77°N. Clearly these coastal regions of high precipitation correspond to regions where there is open water offshore for much of the

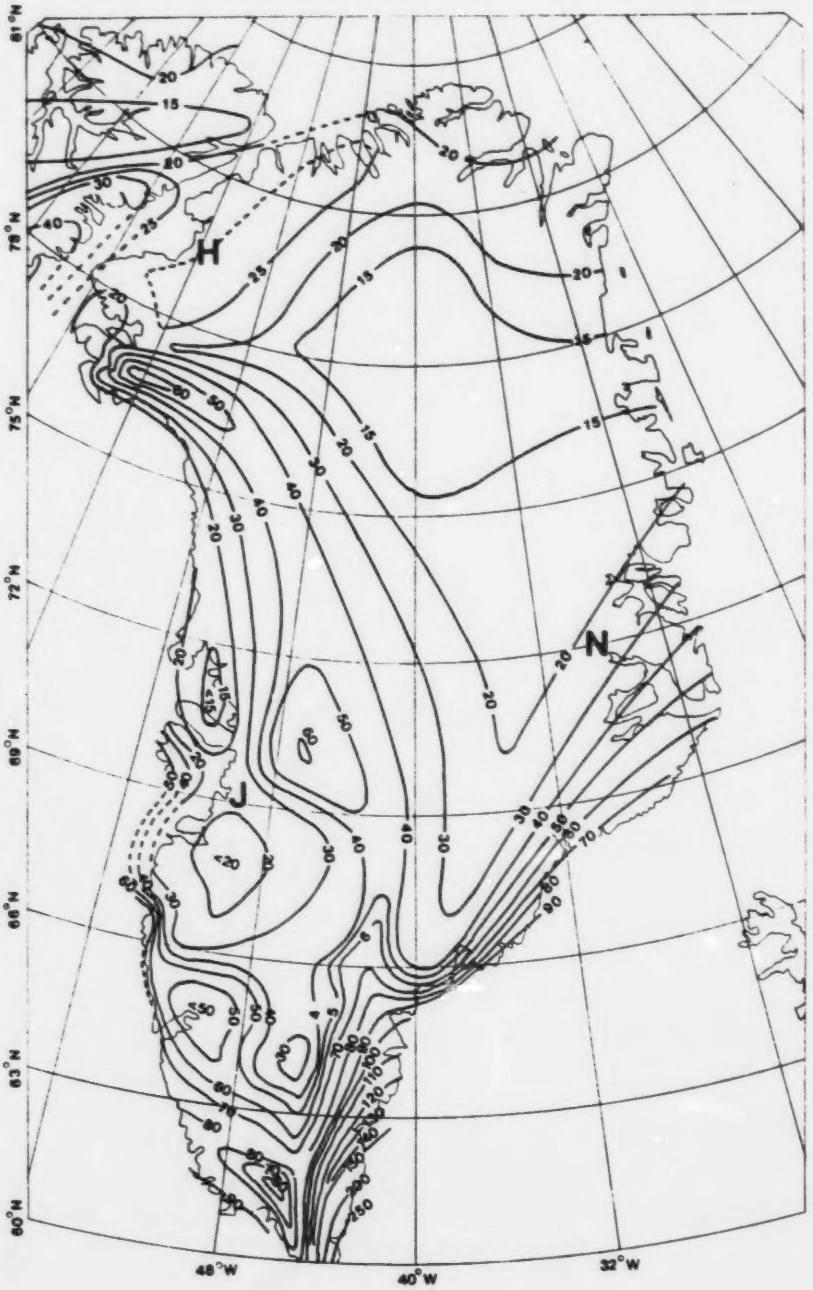

81°N
78°N
75°N
72°N
69°N
66°N
63°N
60°N
48°W
40°W
32°W
H
J
N

Figure B1 Map of estimated distribution of present-day precipitation over Greenland (g/cm^2/yr). The map is based on ice sheet accumulation rate data from snow stratigraphic studies and on precipitation data from coastal stations. (From Reeh, 1985). J is Jakobshavns Isbrae, H is Humboldt Geltscher, and N is Nordvest Gletscher.

year, with the Melville Bay anomaly being a result of the presence of the North Water winter polynya (Ito, 1982). This allows high evaporation and increased regional snowfall. The contrast with coastline which is completely icebound in winter is significant. On the east coast there are 15-20 cm of snowfall (water equivalent) in NE Greenland compared with 40-90 cm in the zone from Scoresby Sund to Angmagssalik, and 90-250 cm southward to Cape Farewell. On the west coast the variation is from about 20 cm in the icebound 70-75°N zone, to 50-60 cm from Disko Island southward and 80-90 cm near Cape Farewell.

In attempting to predict the retreat of the West and East Greenland ice packs due to a specified temperature rise. Our difficulty is that ice dynamics and thermodynamics are both important in determining ice limits. Fast ice is a much easier case, since it occurs in nearshore regions where the oceanic heat flux is close to zero so that its growth rate is almost entirely determined by air temperature. Successful empirical methods of predicting fast ice growth and decay have therefore been developed based on the concept of degree-days of cold (Bilello, 1961) or of thawing (Bilello, 1980). In this method the thickness reached by the ice is related to the sum of the temperatures in °C below -1.8°C achieved on each day of the winter up to the day in question. Wadhams (1989) showed how this technique can be used to predict a 59-day increase in the open water season at Resolute with a 40-70 cm drop in maximum winter ice thickness if an 8°C rise in mean air temperature were to occur. Only one ice thickness dataset exists from the pack ice in Baffin Bay and Davis Strait; Wadhams et al. (1985) obtained the ice thickness distribution in Davis Strait in mid-February 1967 from upward sonar data from a submarine, finding an overall mean thickness for undeformed first-year ice of 1.18 m. Neglecting oceanic heat flux, this would correspond to the effect of some 1800 degree-days of cold: a 5°C temperature rise would reduce the thickness to about 70 cm, while a 13°C rise would be needed to prevent ice formation altogether. Of course, ocean heat flux is highly significant in Davis Strait, and greenhouse warming would undoubtedly be accompanied by a rise in the ocean temperature below the mixed layer with a consequent rise in oceanic heat flux; the sea ice would then disappear with a lower air temperature rise than 13°C.

The best dataset for recent ice limits in Baffin Bay is the passive microwave data from ESMR and SMMR. This was analysed by Parkinson and Cavalieri (1989) for ice extent over the 1973-1987 period, giving minima and maxima of 1.31 and 1.74 million sq km for winter ice extent. This does not tell us where the ice reached along the west Greenland coast, however. These data come from maps presented by Comiso and Zwally (1989) for the years 1978-1987. They show in winter a tongue of open water stretching up the west Greenland coast. The lowest latitude reached by the ice near the coast was 61° (in 1984) and the highest 67° (in 1979). This gives a 6° range over 10 years. We surmise that greenhouse warming may lead to a further 6° retreat relative to the recent warmest year, i.e. give a winter maximum extent at 73°N. This suggests that we should move the zone of high precipitation on the west coast of Greenland northward by 6°.

Baffin Bay forms an almost closed system containing mainly first-year ice, so although the ice is in motion and thus generating ridges and leads, we can to some extent think of thermodynamic growth rates as being relevant to the final quantity of ice seen in Baffin Bay. On the east coast of Greenland, however, the East Greenland pack ice forms an extension of the pack ice regime of the Arctic basin. It has already been shown by tests using the numerical model of Hibler (1980) that the ice limits in the Arctic Ocean are controlled much more by ice dynamics than by ice thermodynamics. Therefore even if a significant warming occurs, it is by no means certain that the ice will retreat, as compensatory changes in wind fields and thus the balance of ice convergence versus divergence may occur. The rate of lead opening in winter is particularly important, as it is here that the most rapid ice growth occurs. To estimate the retreat of sea ice that corresponds to CO_2 doubling we can, as in the case of Baffin Bay, consider the climatic extremes over the last 20-30 years and assume as a first approximation that the additional amelioration of ice conditions (i.e. retreat of the ice edge) will be similar in magnitude to the difference between the best and worst ice year in that past period. Thus we take a natural measure of climatic variability as being of the same order as the deterministic variation due to a distinct warming.

A justification for this technique comes from the modelling work of Herman and Johnson (1978). They ran the Goddard global climatic model with two sets of ice limits as inputs, representing the maximum and minimum winter ice extents over the period 1961 to 1977, but with other input parameters unchanged. They found, for instance, that a minimal ice extent in Davis Strait corresponded to a mean surface temperature rise of 3-4.5°C, which is similar to the magnitude expected from greenhouse warming. By forcing a climate model with a reduced ice extent they obtain a given temperature rise. It is not necessarily true that by forcing a model with a given temperature rise the output will be a similar ice retreat; existing models do not operate in this direction. Nevertheless, it gives some support to the idea that the expected further retreat will be of roughly the same magnitude as the amplitude of natural fluctuations over the last 20 years or so.

In East Greenland a 10-year dataset of extreme fluctuations (1966-75) was presented by Vinje (1977; see also Wadhams, 1986). In winter the maximum ice limit extends around Cape Farewell while the minimum is still south of Denmark Strait. In summer Scoresby Sund represents the minimum extent except for an anomalous retreat in July up to 77°N; the band of ice along the coast is very narrow in summer, however. We surmise that the further retreat due to greenhouse warming might be sufficient to make 77°N the typical ice limit as far as an effect on precipitation is concerned. The data of Comiso and Zwally (1989) also fit these results.

If we extend the coastal zones of heavy precipitation as indicated above, and raise overall precipitations by 10% we get a contour map which may be roughly represented by Figure 2. Other effects may occur: for instance, the low precipitation over the central dome of the Greenland ice cap, extending to the NE, is partly a function of a ridge of high pressure which exists over NE Greenland in long-term average pressure maps (e.g. Parkinson et al., 1987). Sea ice retreat in the Greenland Sea would cause this ridge to retreat, which may cause a more radical rise in precipitation over the central Greenland ice cap.

2.2. Equilibrium line.

The two scenarios for the rise in equilibrium line were selected by Bindschadler (1985) because of a large uncertainty in the temperature gradient with altitude on the Greenland ice sheet. He used the prediction of Hansen et al. (1984) that a CO_2 doubling will lead to a rise in mean surface temperature of 4°C in southern Greenland and 6°C in northern Greenland. Two widely differing values are available for the temperature gradient with altitude: 0.6°C per 100 m for the EGIG line at 70-71°N (Ambach, 1982); and 1.12°C per 100 m for Camp Century at 76.5°N (Benson, 1962). The two scenarios of a 500 m and 1000 m rise in the equilibrium line altitude bracket the possible values that can be derived from the two altitude gradients and the range of surface temperature rises.

2.3. Basis of Modelling.

Bindschadler (1985) applied these climatic changes to a simple two-dimensional static model of the entire Greenland ice sheet, giving a shape for the ice cap and a relation between central elevation and width. He calculated a static response to climatic change, obtaining a large negative mass balance; the most interesting aspect of this is the finding that the loss of accumulation area due to the rise in the equilibrium line more than offsets the increase in accumulation rate due to the rise in precipitation. However, this model cannot deal with calving, and since the model is a static one based on shape alone the negative mass balance manifests itself as a retreat and thinning of the ice front region.

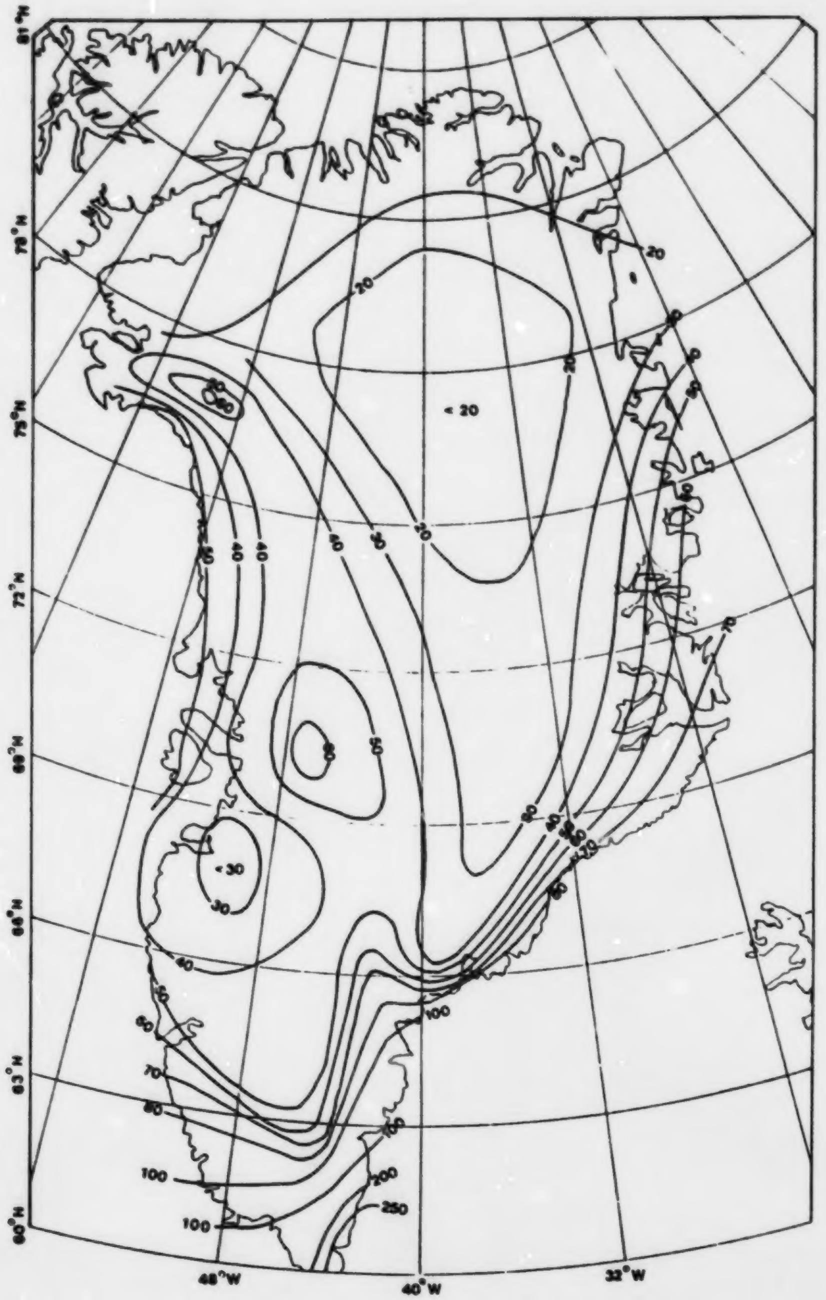

81°N
78°N
75°N
72°N
69°N
66°N
63°N
60°N
20
< 20
30
< 30
40
50
60
70
80
100
200
250
48°W
40°W
32°W

Figure B2 Map of estimated climate-induced 10% overall increase in precipitation over Greenland, modified by the effects of changes in the distribution of sea ice.

In considering a possible change in calving rate, Bindschadler mentions the work of Sikonia (1982; see also Brown et al., 1982a), who found an empirical relationship for the calving speed of Columbia Glacier in Alaska, of the form

$$V_s = 1.092 \times 10^3 D^{0.5689} h^{2.175} \qquad (1)$$

where V_s is calving speed in km/yr, D is water discharge rate in m^3/s and h is unsupported height of the ice column in the water. Brown et al. gave more approximate exponents of 0.5 and 2. Bindschadler considered that the increased meltwater discharge from the Greenland Ice Sheet will encourage iceberg calving in the same way, due to the water discharge speeding up the flow rate of the ice sheet. He estimated that calving rates will increase to a factor of 1.78 (scenario A) or 3.07 (scenario B) times present levels due to this effect. A problem with the application of equation 1 to the case of many fast-flowing ice sheet outlets (e.g. Jakobshavns Isbrae, Nordvest Gletscher) is that their termini are floating, whereas that of Columbia Glacier is grounded. It is unlikely, therefore, that the constants derived from the latter case are applicable to calving rates from ice masses with floating margins.

The method which we have adopted in this report is to use calculations from a balance flux model, of the kind used by Bindschadler (1984) and Reeh (1985). The balance or ice flux at the lower, tidewater end of any ice drainage basin is equal to the rate of iceberg production. The calculation is applied to the Greenland Ice Sheet to estimate present iceberg production and climate-induced changes in that flux. The results of these calculations are then compared with other measurements and calculations of present ice production, and predictions of future changes. The problem in this, as in all previous, analyses is the lack of observational data from very large areas of the Greenland Ice Sheet. This renders more complex modelling problematic, because the boundary conditions required to constrain the analysis are largely unavailable. The drainage basin with which we will deal in detail is that of Jakobshavns Isbrae in West Greenland at 69°N. This is the only major basin for which reasonably detailed data are available in the published literature.

3. CALCULATIONS OF ICEBERG PRODUCTION

3.1. Data Sources

A major constraint on any calculations of the response of the Greenland Ice Sheet, and its production of icebergs, in a changed climate is the lack of observational data available even for the present. This severely restricts the areas of the ice sheet for which meaningful analyses can be undertaken. There is at present no general agreement even about what is the *modern* rate of ice flux into the marine environment from the whole ice sheet.

Jakobshavns Isbrae (69°N)

The most detailed data coverage for any drainage basin of the Greenland Ice Sheet is that for Jakobshavns Isbrae (Fig. 1), a fast-flowing outlet glacier with a long profile from terminus to ice divide of some 570 km (Bindschadler, 1984). Observations in or adjacent to this drainage basin include:

(a) Mass balance from the EGIG Traverse in 1959 (Benson, 1962)
(b) Ice surface elevations from the SEASAT satellite radar altimeter (Bindschadler, 1984)
(c) Iceberg production rate and terminus velocity from observations, calculations and aerial photographs (Carbonnell and Bauer, 1968; Lingle et al., 1981; Bindschadler, 1984; Echelmeyer, pers. comm.)
(d) Ice surface velocities in the upper basin from the EGIG Traverse (Mälzer, 1964)
(e) Ice thickness compiled from radar measurements (Radok et al., 1982)

Of particular importance to our calculations of ice flux are the measures of drainage basin area and mass balance (accumulation and ablation). Drainage basin area is particularly problematic because very low slope angles (<<1°) in the upper basin mean that minimum and maximum basin area estimates vary by over 30%. The Jakobshavns basin is covered by SEASAT radar altimeter data accurate to 1-2 m (Zwally et al., 1983), but even here problems occur over the spatial scale at which surface slope is determined. We have taken the minimum and maximum estimates of basin area from Bindschadler's (1984) analysis of SEASAT data. This satellite analysis of basin dimensions is a considerable improvement over previous maps, which were constrained by the very limited availability of elevation data (Weidick and Olesen, 1978; Weidick, 1980).

Measurements of mass accumulation in the basin are taken from the observations of accumulation changes with elevation by Benson (1962). Reeh (1985) presents a version of Benson's compilation for the whole of the Greenland Ice Sheet, which has been modified to include more recent data especially for the southern part of the Ice Sheet (Fig. 1). In particular the accumulation rate in southwest Greenland between 65 and 69°N has been revised downwards relative to previous estimates (e.g. Mock, 1967). The elevation of the equilibrium line is also derived from observations on changes in the surface characteristics of the ice by Benson during the EGIG Traverse. It is placed at 1,300 m above sea level. This is the value we use here, although Ambach (1982) places it at 1,500 m. Below this level ablation exceeds accumulation through a balance year. Bindschadler (1985) provides an empirical equation for the calculation of ablation rate with elevation using a linear relation between increasing ablation and decreasing elevation:

$$b = db/dh\ (H_e - h) \qquad (2)$$

where b is ablation rate, H_e is equilibrium line altitude, and h is elevation at a given point on the ice cap. Braithwaite (1980) has measurements of ablation from 72°N, where db/dh is 0.00153/yr. This value is used in our ablation calculations.

Other Greenland Drainage Basins

At the outset of the study it was intended to undertake a similar analysis to that for Jakobshavn Isbrae for two other large drainage basins which produce significant fluxes of ice to the oceans (Fig. 1). The first was Humboldt Gletscher (79-80°N in NW Greenland), whose 111 km-wide terminus is a major iceberg source (Feazel and Kollmeyer, 1972; Kollmeyer, 1980). The second was Nordvest Gletscher (72°N); a fast-flowing outlet glacier entering inner Scoresby Sund in East Greenland (Olesen and Reeh, 1969). This would have given a spatial coverage around the Greenland Ice Sheet, and allowed comparison of any differences in mass balance regime and basin hypsometry to be investigated. Reeh (1985) estimates that Humboldt and Nordvest Gletscher today produce approximately 7 and 11 km^3/yr of icebergs, respectively, out of a total Greenland production of approximately 310 km^3/yr.

However, attempting a similar form of balance flux analysis for any other major drainage basin of the Greenland Ice Sheet is problematic, because of the lack of data specifying even *contemporary* basin area and mass inputs. SEASAT altimetric data are not available north of 72°N (Zwally et al., 1983), making the definition of the upper basins within the Ice Sheet north of this area difficult. Mass balance data are restricted to the paths of traverses of Greenland and to sites of deep drilling and other detailed studies. The errors in basin area alone for Jakobshavns Isbrae induce an error envelope of up to 18 km^3/yr or over 35% in the calculation of present ice flux (Bindschadler, 1984). In basins for which the area is less well defined (i.e. all other large basins on the Ice Sheet) a detailed analysis of the type carried out for Jakobshavns loses most of its value.

3.2. Method of Balance Flux Calculation

The method of calculating the balance flux across a transverse section of an ice drainage basin follows that of Bindschadler (1984) and relies on the assumption of mass continuity. A similar approach has been taken by Reeh (1985) to the estimation of iceberg production from the entire tidewater margin of the Greenland Ice Sheet (Fig. 3). Reeh points out that a reliable estimate of contemporary ice production from the whole Ice Sheet is not possible at present due to insufficient observational coverage. He states that the lack of data is worst in northwest, north and east Greenland.

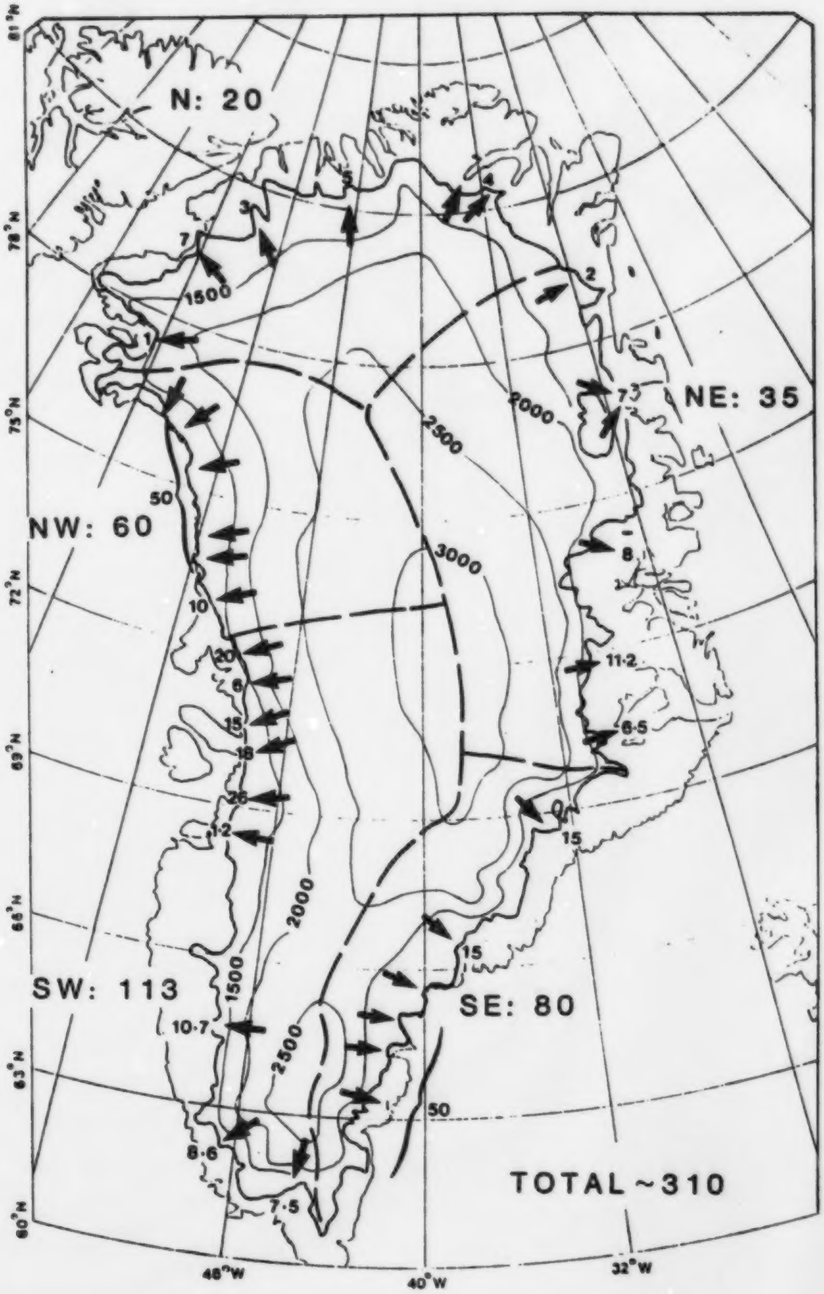
The balance flux at a cross section equals the areal integral of net mass balance over the area upstream of the cross section:

$$Q(x) = \int_0^x W(\xi)\, b(\xi)\, d(\xi) \qquad (3)$$

where Q is balance flux at location x along glacier long profile, W is width across a basin, and b is net balance averaged across the width. Note that balance flux, required to provide estimates of ice volume discharged to the sea, can be calculated without a knowledge of ice depth or the 3-dimensional shape of the basin. Only surface measurements are required.

The balance or ice volume flux at the tidewater margin of an ice drainage basin gives values for the production of icebergs. Varying the mass balance term in equation 3 allows us to assess the changes in ice flux associated with the two climatic change senarios given in Section 1. For Jakobshavns Isbrae the relationships between mass balance and elevation used for calculations are given in Figure 4 and Tables 1 and 2. The curve for present mass balance is a modified version of that used by Bindschadler (1984). In our analysis, which places the modern equilibrium line at 1,300 m (Benson, 1962), we have included a net mass loss below this elevation (Fig. 4). The other two curves are for the two climate change senarios specified in Section 1. The rate of ablation below the upward-shifted equilibrium line is calculated from the empirical relationship in equation 2. Above the equilibrium line the shape of the curve for net accumulation is assumed to remain as for the present environment, but it is displaced up-glacier. A 10% increase in precipitation for the two climate change scenarios is also included in these two curves. We have not included the estimated changes in near-coast precipitation related to climate-induced sea ice retreat up Baffin Bay (Fig. 2). This is because such a small proportion of total basin area lies close to the coast (Fig. 5) that the effect of this change would be insignificant.

The horizontal extent of the Jakobshavns drainage basin is assumed to remain constant in all calculations. Ice fluxes are calculated using both the maximum and minimum estimates of basin size provided by Bindschadler (1984). Hypsometric curves for both maximum and minimum basins are shown in Figure 5. Note that 500 m and 1,000 m rises in ELA are equivalent to distances from the glacier terminus of 150 km and 250 km, respectively. It is essentially the combination of data in Figures 4 and 5 that yields our ice flux values when applied to equation 3.


81°N
78°N
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63°N
60°N
48°W
40°W
32°W
N: 20
NE: 35
NW: 60
SW: 113
SE: 80
TOTAL ~310
1500
2000
2500
3000
2000
1500
2500
3
7
5
4
2
1
7·3
8
11·2
6·5
50
10
20
6
15
18
26
1·2
10·7
8·6
7·5
15
15
1
50

Figure B3 Estimated present-day iceberg production from the Greenland Ice Sheet. The values given are for major outlet glaciers and the total annual production for each of several geographical regions of the Ice Sheet. Values are km^3/yr of water equivalent. (From Reeh, 1985).

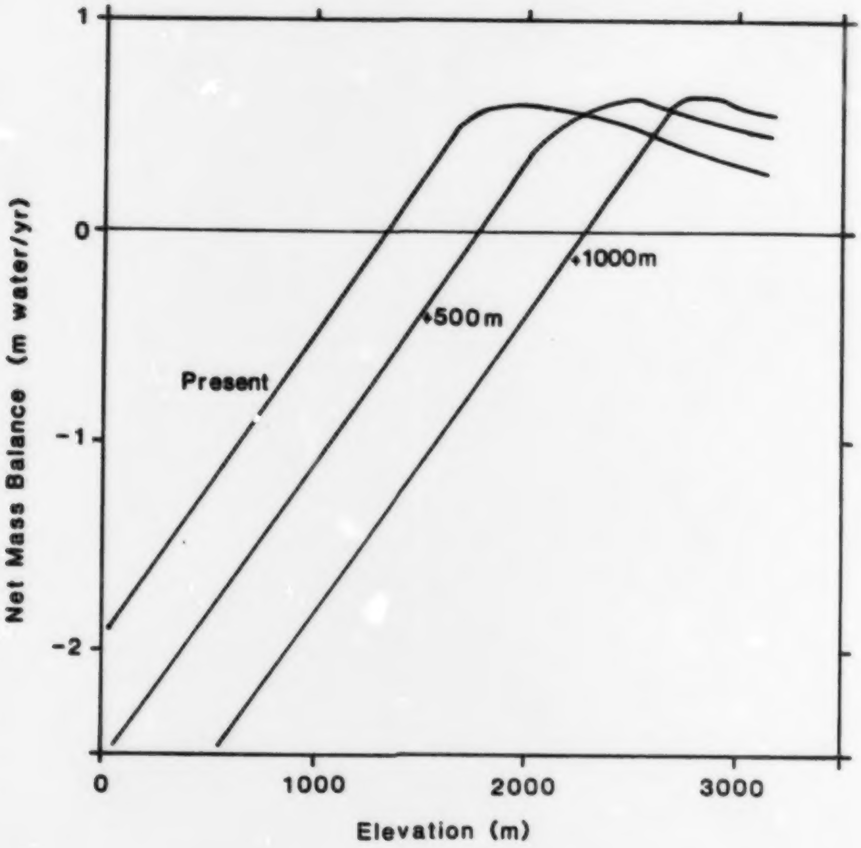

Figure B4 The relationship between mass balance and elevation for the Jakobshavns Isbrae basin. The curves represent the estimated contemporary relationship and those relating to the equilibrium line altitude shifts specified in climate change Scenarios A and B.

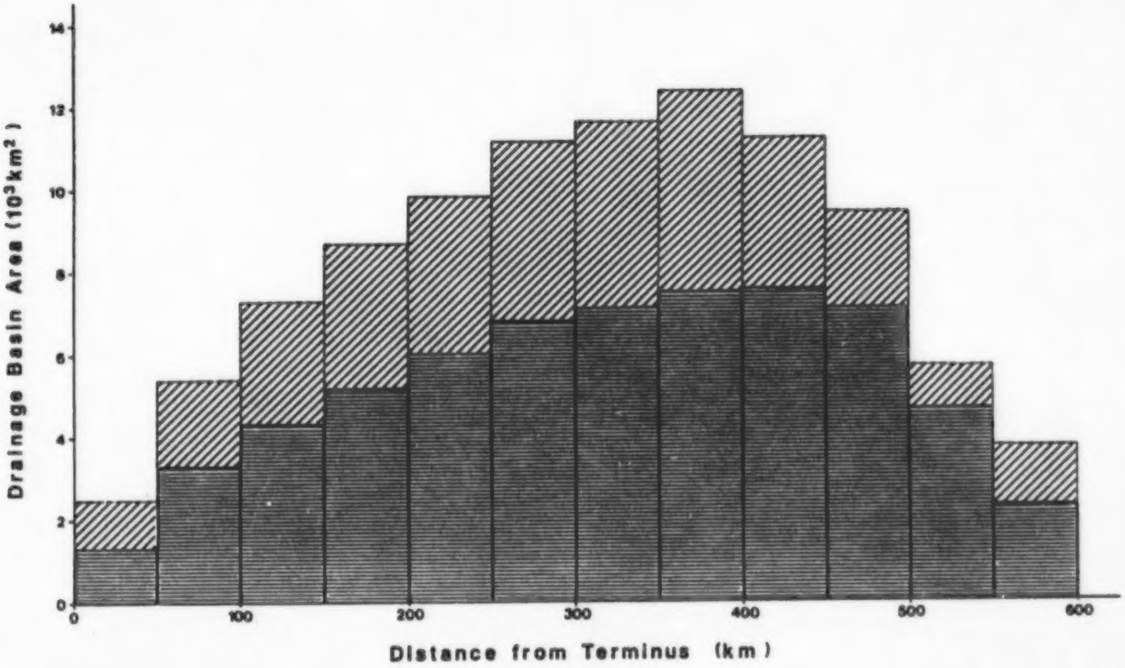

Figure B5 Minimum and maximum estimates of the hypsometric curve for Jakobshavns Isbrae basin. (Data from Bindschadler, 1984).

3.3. Results

The results of the calculations of balance or ice flux for Jakobshavns Isbrae are given in Tables 1 and 2. Table 1 contains the results of our calculations for the two cases of changed climate: 500 m rise in ELA plus 10% precipitation rise, and 1,000 m rise in ELA plus 10% precipitation increase. Table 2 provides values of ice flux for the present climatic conditions. It is the comparison of the ice fluxes in these tables that provides an estimate of the change in iceberg production from the Jakobshavns basin due to greenhouse warming.

Values of balance or ice flux at the terminus of Jakobshavns Isbrae are of principal interest in this study, since this is effectively the volume of iceberg production. Bindschadler (1984) calculates present flux to be between 30 and 48 km^3/yr. Our revised calculation, including ablation below 1,300 m, yields the lower values of 27-42 km^3/yr. Applying the more modest of the two climate change scenarios to the ice flux calculation gives values between 25 and 38 km^3/yr. When data for the more extreme climate warming scenario are used, ice flux at the margin drops to between 14 and 17 km^3/yr. The results of this analysis suggest that only in the case of a 1,000 m rise in equilibrium line altitude will climate change be severe enough to induce a significant change in ice production at the terminus of Jakobshavns Isbrae. The interpretation of these finding is given below.

3.4. Discussion and Comparison with Other Work for Jakobshavns

Present Iceberg Production. A comparison between our own calculations and several other studies of ice production from the Jakobshavns basin is given in Figure 6. The estimates of present-day iceberg yield by Carbonnell and Bauer (1968) and calculated from Lingle et al. (1981) are based on analyses of aerial photographs and direct survey data. Echelmeyer (pers. comm.) also provides new survey and ice thickness data from the ongoing prject on Jakobshavns terminus yielding a rate of iceberg production of 42 km^3/yr. Each study is independent of the method used by ourselves and Bindschadler (1984) for calculating contemporary fluxes. It can be seen that the Bindschadler, Carbonnell and Bauer, Echelmeyer and Lingle et al.-derived values overlap and, within the considerable errors in the methods used, are similar. Our downward revision of Bindschadler's values (to 27-42 km^3/yr) also falls within the same envelope. Thus, we can be fairly confident that present iceberg production from Jakobshavns Isbrae ranges between 28 and 48 km^3/yr.

Scenario A (500 m rise in ELA plus 10% increase in accumulation). Our calculations for the less severe climatic warming of Scenario A yield a value of iceberg

TABLE 1a. Calculations of balance flux for Jakobshavns Isbrae based on a 10% increase in precipitation and 500 m rise in equilibrium line altitude. Balance flux for the 0-50 km segment represents iceberg production.

Distance from Terminus (km)	Net Balance (m water/yr)	Area (10^3 km^2)		Bal. flux (km^3 ice/yr)	
		Min.	Max.	Min.	Max.
0-50	-1.51	1.32	2.50	24.99	37.51
50-100	-0.70	3.28	5.39	27.19	41.61
100-150	-0.13	4.30	7.27	29.71	45.76
150-200	0.36	5.16	8.68	30.32	46.80
200-250	0.51	6.02	9.85	28.46	43.68
250-300	0.63	6.80	11.18	25.39	38.66
300-350	0.59	7.11	11.64	20.68	30.92
350-400	0.55	7.50	12.19	16.07	23.37
400-450	0.52	7.58	11.25	11.54	16.00
450-500	0.48	7.11	9.45	7.21	9.57
500-550	0.45	4.84	5.70	3.46	4.59
550-570	0.43	2.27	3.75	1.07	1.77

TABLE 1b. Calculations of balance flux for Jakobshavns Isbrae based on a 10% increase in precipitation and 1000 m rise in equilibrium line altitude. Balance flux for the 0-50 km segment represents iceberg production.

Distance from Terminus (km)	Net Balance (m water/yr)	Area (10^3 km^2)		Bal. flux (km^3 ice/yr)	
		Min.	Max.	Min.	Max.
0-50	-2.20	1.32	2.50	13.78	17.81
50-100	-1.38	3.28	5.39	16.97	23.85
100-150	-0.83	4.30	7.27	21.94	32.02
150-200	-0.41	5.16	8.68	25.19	37.51
200-250	-0.07	6.02	9.85	27.52	41.42
250-300	0.51	6.80	11.18	27.98	42.18
300-350	0.57	7.11	11.64	24.17	35.91
350-400	0.65	7.50	12.19	19.72	28.62
400-450	0.64	7.58	11.25	14.36	19.91
450-500	0.59	7.11	9.45	9.03	12.00
500-550	0.58	4.84	5.70	4.50	5.98
550-570	0.57	2.27	3.75	1.42	2.35

TABLE 2. Calculations of the present balance flux for Jakobshavns Isbrae. Values are those of Bindschadler (1984) except for those between 0 and 100 km from the terminus, where we include net ablation below 1,300 m in elevation. Balance flux for the 0-50 km segment represents iceberg production.

Distance from Terminus (km)	Net Balance (m water/yr)	Area (10^3 km^2)		Bal. flux (km^3 ice/yr)	
		Min.	Max.	Min.	Max.
0-50	-0.90	1.32	2.50	26.98	42.11
50-100	0.00	3.28	5.39	28.30	44.61
100-150	0.55	4.30	7.27	28.30	44.61
150-200	0.59	5.16	8.68	25.67	40.17
200-250	0.55	6.02	9.85	22.30	34.49
250-300	0.50	6.80	11.18	18.62	28.48
300-350	0.45	7.11	11.64	14.85	22.27
350-400	0.40	7.50	12.19	11.29	16.45
400-450	0.36	7.58	11.25	7.96	11.03
450-500	0.33	7.11	9.45	4.93	6.53
500-550	0.30	4.84	5.70	2.32	3.07
550-570	0.28	2.27	3.75	0.70	1.17

production between 25 and 38 km^3/yr (Fig. 6; Table 1a). This is not significantly different from the envelope of present-day estimates of ice production. The reason for this is that, although ablation on the lower glacier is increased (Fig. 4), the peak in net accumulation values is: (i) enhanced by the specified 10% increase in precipitation; and (ii) shifted up-glacier towards the widest part of the drainage basin (Fig. 5). Thus, increased precipitation falls over a larger area of the catchment, increasing balance flux from the central part of the basin. The effect of this is to offset the trend to greater ablation in the lower glacier and to yield a relatively unchanged flux of icebergs.

Scenario B (1,000 m rise in ELA plus 10% increase in accumulation). In the greater warming associated with Scenario B, the up-glacier shift of the ELA, and the further enhanced ablation associated with it (Fig. 4), can no longer be offset by the 10% precipitation increase and the fact that basin area and accumulation are weighted towards the upper basin. The comparisons in Figure 6 indicate that ice flux at the basin margins under Scenario B is only just over 50% of its present value or that associated with Scenario A. Thus, the calculations suggest that only in the case of the more severe climatic warming scenario is iceberg production cut significantly. However, the assumptions on which each of the values presented in Figure 6 is based should be remembered in assessing this statement.

In particular, it should be noted that the balance flux calculations do not take account of any lag time between changes in precipitation and ablation rates and the propagation of that effect down-stream. At the velocities of less than 50-100 m/yr observed in the upper part of the Jakobshavns basin by Mälzer (1964), it will take several thousand years for ice build-up in the upper basin to be reflected in changes at the coast (cf. Radok et al., 1982, Figure 2/9). This assumes that kinematic wave behaviour does not take place (cf. Nye, 1960, 1963). Detailed modelling of such effects is beyond the scope of this report.

4. IMPLICATIONS FOR ICEBERG PRODUCTION FROM THE GREENLAND ICE SHEET

4.1. The Direction and Magnitude of Change

Our calculations indicate that, for Jakobshavns Isbrae, iceberg production will remain similar to that of today under climate change Scenario A, but be curtailed by some 50% with more severe climatic warming. The time scale on which these changes operate is uncertain due to lags in ice sheet response to changed mass inputs. The direction of change in iceberg production appears, therefore, to be towards a decrease in iceberg calving for the more major climate change specified. The calculations we

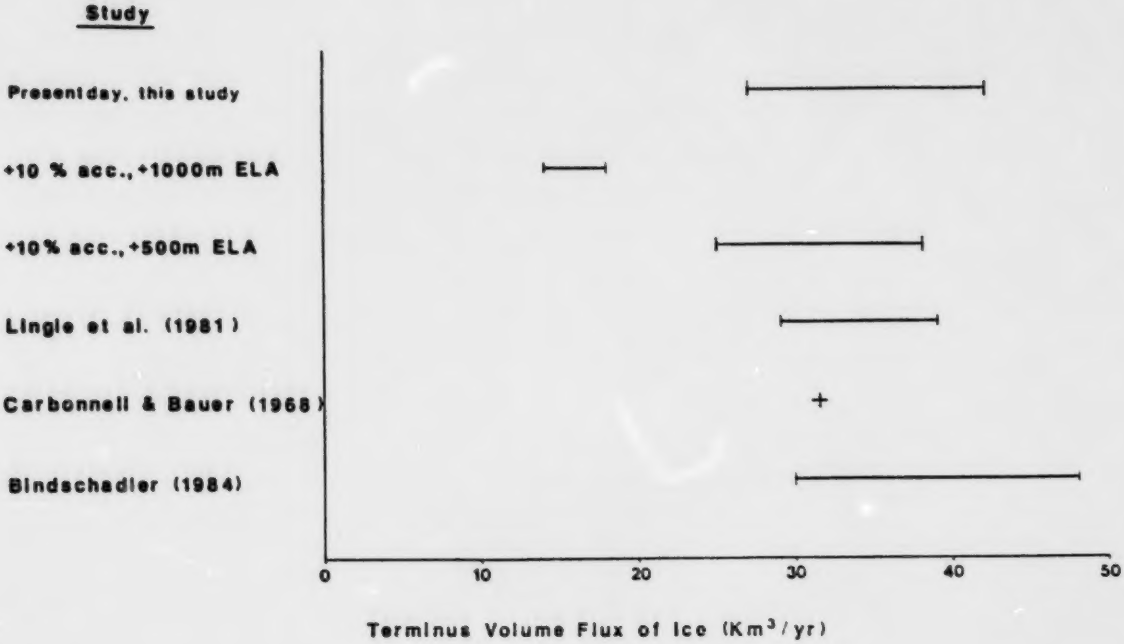

Figure B6 Observed and calculated values for the volume flux of ice at the terminus of Jakobshavns Isbrae. Values from Bindschadler, Carbonnell and Bauer, and Lingle et al. are estimated present-day fluxes. The other values are for climate change Scenarios A and B.

have made do not indicate an increase in iceberg calving for either climate change scenario. However, transient ice flow effects and local outlet glacier responses (Section 5) are not included here, and the possibility that they may lead to short-lived increases in iceberg production cannot be ruled out.

Reeh (1985) suggests that total present iceberg production from the Greenland Ice Sheet is approximately 310 km^3/yr of water equivalent or 344 km^3/yr of ice equivalent in our terms (Fig. 3). He points out, however, that only 45% of this value is based on observations of glacier terminus velocities and thicknesses. Data from the northwest, north and east of Greenland are few, and he states that his overall value should be treated with considerable caution. Reeh's present-day ice production estimate for Jakobshavns Isbrae is 26 km^3/yr. Our calculation for ice production under climate change Senario B suggests a reduction of about 40-50% from this value. If a similar climate-induced change were to occur over the whole ice sheet, iceberg production would be cut by some 140-170 km^3/yr to about 170-200 km^3/yr. Clearly, there are very real dangers in extrapolating this magnitude of change to the whole ice sheet, and it really requires detailed studies in several large drainage basins spead over the ice sheet in order to come to more specific conclusions.

4.2. Comparisons with Other Studies

The aim of most workers has been to assess the implications for global sea level of climate-induced changes in the configuration of the Greenland Ice Sheet rather than to consider the direct effects of warming on iceberg production. However, the most closely related study to our own is that of Bindschadler (1985). He also specified two future climatic changes of 500 m and 1,000 m rises in ELA, but did not include the 10% accumulation increases used in our calculations. His two-dimensional flow-line model has been described in Section 2.2. He suggests that under both climate changes the ice sheet will experience overall mass loss: 460 km^3/yr for a 500 m ELA rise and 1,260 km^3/yr for a 1,000 m ELA rise. He does not applied these values directly to a revising of iceberg production rates, because his model holds this parameter constant thoughout the analysis.

Recent analyses and intercomparison of satellite radar altimeter data from GEOS-3, SEASAT and GEOSAT have demonstrated that the Greenland Ice Sheet to the limit of data at 72°N has thickened since the mid-1970s (Zwally et al., 1989).The thickening is of approximately 0.23 m/yr. This, according to Zwally (1989), suggests that precipitation, at least over the southern 40% of the ice sheet, is higher than the existing long term average. This may be an observable result of climate warming that has already occurred in the Arctic (Zwally, 1989). Comparison between balance flux calculations for climate change Scenarios A and B (Table 1) with the present-day situation (Table 2) indicates that mass inputs to the upper basin will rise relative to present levels. The findings of Zwally indicate the same kind of trend as our calculations, and the observed thickening is presumably taking place because ice flow does not adjust instantaneously to changing mass inputs.

5. SOURCES OF UNCERTAINTY

5.1. Transient Response

This analysis has assumed that the climatic changes are slow enough for the flow of the ice sheet to adjust so as to remain in quasi-equilibrium with the new forcing functions. However, a dynamic response of a glacier also exists. Nye (1960, 1963) showed that if a change in accumulation rate is imposed on all or part of a glacier the effect is initially propagated downstream in the form of a kinematic wave. This piles up against the lower end of the glacier and results in a transient increase in thickness and/or a surge forward of the glacier terminus. Clearly this would lead to a transient increase in iceberg calving rate. Time scales for kinematic wave propagation are in the range 3-30 years, depending on glacier size. Paterson (1981) has pointed out that glaciers are constantly adjusting to local changes and variations in accumulation rate and temperature, and that there are few clear recorded instances of kinematic waves. The result of such a wave, relative to the equilibrium adjustment predicted by the flowline model, would be an initial excess of icebergs, followed by a dearth of icebergs when the additional mass has been discharged. We feel that the time scale of the climatic change considered places it just beyond the point at which it can be considered a step function impact capable of generating a kinematic wave. We therefore do not consider this transient response as being likely to play a major role.

5.2. Local Effects in Fjords

Iceberg discharge does not occur at a uniform rate throughout the year. Calving rates appear to be greatest in spring, when meltwater increases the flow speed of the lower part of the glacier. Icebergs once calved cannot, however, move out of their parent fjords until the fast ice breaks up. There is strong evidence (Dowdeswell, 1989) that fast ice prevents iceberg movement. It has also been observed in the Antarctic (Wadhams, personal observations, 1989) that despite the difference in magnitude and direction of the driving forces on icebergs as opposed to pack ice, icebergs embedded in large sea ice floes move with the floe rather than ploughing a track through the sea ice, even with the sea ice as thin as 60 cm.

Climatic warming will lead to an earlier breakup of fast ice in fjords and a later formation, with the ice-free season being extended by a period of the order of two months (Wadhams, 1989). This will allow icebergs to move out of the fjord more easily. The shorter residence time of bergs in fjords may mean less melt occurring in the fjord (except for bergs initially too deep to pass the fjord sill) although this will be offset by the warmer water temperature. We assume that the impact of these effects on iceberg flux into the West Greenland Current is small enough to be ignored.

Warmer water temperatures within fjords will also result in increased direct melting of the glacier terminal ice cliffs (M_d). Increased direct melting will tend to reduce the rate of iceberg calving (V_c), because the two are related through the equation (Dowdeswell and Murray, in press):

$$V_c = u - m - M_d \tag{4}$$

where u is velocity at the tidewater glacier terminus and m is the rate of advance or retreat of the terminus. We can calculate the response of the term M_d to climate-induced increases in water temperature through the following equation (Weeks and Campbell, 1973):

$$M_d = 6.74 \times 10^{-6} V_w^{0.8} \Delta T / X^{0.2} \tag{5}$$

where V_w is free stream vertical water velocity, X is height of the ice cliff in the water column, and ΔT is the temperature difference between the ice and the surrounding water. The effect of even an extreme rise in water temperature of 6°C is to increase melting back of the ice cliffs by some 40-50 m/yr (Dowdeswell and Murray, in press). This is an insignificant value relative to calving rates of 3-7,000 m/yr for fast-flowing Greenland outlet glaciers (Olesen and Reeh, 1969; Reeh, 1985), and climate-related changes in this parameter can also be ignored.

The topography of the fjord floor beneath calving outlet glaciers can also exert a significant local control on iceberg production. The calving rate of grounded, as opposed to floating, ice fronts has been shown to be strongly dependent on water depth. The ice mass presumably becomes more buoyant and ,therefore, calving become more rapid as depth increases. Brown et al. (1982b) have derived an empirical function for this relationship, based on a study of 14 Alaskan tidewater glaciers,

$$V_c = k\, h_w \tag{6}$$

where V_c is calving rate, h_w is water depth and k is an empirical constant with a value of 27/yr. Thus, if a glacier retreats into a trough in its bed profile, the calving rate will increase greatly for non-climatic reasons. A initial, climate-induced retreat of the terminus could lead to a topographically-induced increase or decrease in calving rate if major variations were present in the glacier bed profile. This is a local effect specific to each outlet glacier of the Ice Sheet, and is not considered in detail here. The likelihood is that random variations in bed topography would even out over a large number of outlet glaciers, and the effects on total iceberg production from Greenland would be limited.

Fjord bathymetry and the nature of currents and drift tracks influence the rate at which icebergs, once calved, leave the parent fjord and enter the West Greenland Current. Both these factors are specific to individual fjords. Landsat satellite data have been used to monitor iceberg drift tracks within Jakobshavns fjord (Birnie and Williams, 1985). Feazel and Kollmeyer (1972) suggest that icebergs may ground, for example, beyond the margins of Humboldt Gletscher in NW Greenland.

5.3. Other Climatic Inputs

We also neglect a number of other climatic change factors which have been postulated as important in the cryosphere. One is a possible reduction in the albedo of Greenland snow due to increasing pollution by industrial soot (Chylek et al, 1983). This would cause an increase in short wave radiation absorption, and act in the opposite direction to the effect of Arctic haze, which results from industrial particulates in the atmosphere (Clarke and Noone, 1985). We have to assume that the temperature and precipitation effects are dominant.

6. CONCLUSIONS AND RECOMMENDATIONS

6.1. Summary of Findings

The findings of balance flux calculations of iceberg production rates for the present-day and for two specified climatic warming scenarios are as follows:

1. Iceberg production calculated for the present using the balance flux approach agrees quite well with empirical observational studies of iceberg calving from Jakobshavns Isbrae, thus suggesting that our approach is an appropriate one.

2. Iceberg production under climate change Scenario A gives no significant change in the volume of ice calved from the Jakobshavns basin - increased precipitation over a larger portion of the basin is offsetting increased ablation.

3. For the more extreme climate change of Scenario B, iceberg production will fall by 40-50%, as increased ablation comes to dominate over a modest precipitation rise on the upper glacier.

It is not clear whether or not other Greenland drainage basins would respond in a similar manner. If we assume that similar changes do take place over the whole ice sheet, iceberg production will be cut by an estimated 140-170 km^3/yr to about 170-200 km^3/yr for Scenario B. However, certain caveats are associated with these conclusions.

Uncertainties concerning (a) the nature of future climate change and (b) the datasets and methods through which calculations of iceberg production are made, imply that the conclusions offered on changing iceberg flux with greenhouse warming must be treated with caution. Taking up the first of these points, two climatic change scenarios were specified at the outset of this study. However, both the exact nature of climatic response to increasing levels of atmospheric greenhouse gases, and the coupling between sea ice retreat, temperature and precipitation changes, are not clearly understood. For example, we have assumed that a 10% precipitation rise is associated with both of the warming scenarios, whereas it appears possible that the increase in accumulation will be greater than this, especially in coastal regions, with some evidence of a change already in progress (Zwally, 1989; Zwally et al., 1989). On the second point, there are uncertainties in the method of calculating iceberg production linked to both the very limited amount of observational data available from the Greenland Ice Sheet and to the neglect of any short term, transient response of the ice mass to changes in its mass balance.

6.2. Recommendations for Further Studies

A problem with any modelling study aimed at predicting future changes in iceberg production is that observational data on boundary conditions and assumed changes in their values through time are both scarce and limiting. We recommend that futher studies of iceberg production from the Greenland Ice Sheet take a more direct approach to the problem, which avoids many of the dubious assumptions required in modelling work. This approach involves the *direct monitoring* of key parameters associated with ice sheet surface elevation changes and iceberg production *through time*. These parameters are:

1. Accumulation over the ice sheet.
2. Iceberg calving rates from West Greenland tidewater ice fronts.
3. The westward flux of icebergs derived from East Greenland around Cape Farewell.

These observational data comprise the major inputs and outputs of mass to and from the Greenland Ice Sheet. Each can be monitored though time using *remote sensing methods*, and is discussed in turn below.

1. The use of satellite radar altimeter data by Zwally (1989) and Zwally et al. (1989) to measure changes in the elevation (i.e. the thickening or thinning) of the ice sheet has been described above (Section 4.2). The forthcoming launch of ERS-1 (1991) into an 82° orbit will allow radar altimeter data to be acquired over the whole of the Greenland Ice Sheet, including the part north of the 72° orbit of SEASAT. Algorithms for the

analysis of these data and expertise in their interpretation are already available. Time series radar altimeter data will be analysed as part of the European Space Agency Polar Ice Sheets Programme (PISP), in which the authors are participants. Data on the rate of surface change through time for the Greenland Ice Sheet as a whole should, therefore, become available over the next two to three years.

2. The direct measurement of iceberg production from tidewater termini of Greenland Ice Sheet outlet glaciers can be achieved through analysis of near-terminus velocities derived from sequential SPOT or Landsat visible and near-infrared imagery and from ERS-1 (and later Radarsat) synthetic aperture radar (SAR). We suggest a further programme involving two elements: (a) detailed analysis of sequential, digital satellite data from SPOT, Landsat and SAR sensors for 2-3 major outlet glaciers from different drainage basins; for example, the Humboldt, Upernavik and Jakobshavns glaciers. Detailed analysis would show the appropriate spatial resolution and repeat coverage required to derive surface velocity data and terminal advance/retreat rates needed to calculate iceberg production using equation 4 (Section 5.2). (b) Coverage of the whole West Greenland margin at an appropriate repeat interval (probably annually) to allow direct measurements of iceberg production each year. If, for example, Landsat digital data were found to be of suitable resolution, then approximately 10 scenes would be needed to cover the whole coast at a given time interval. Repeating the analysis each year over a number of years would yield a time series of unequivocal observations revealing any changes in iceberg production.

3. The monitoring of iceberg flux in the narrow (40-50 km wide) westward flowing coastal current around Cape Farewell means that a detailed analysis of iceberg production is unnecessary for East Greenland. Measurements at Cape Farewell integrate the effects of total East Greenland iceberg production and any melting and disintegration that has taken place during travel down the east coast. Remotely-sensed data would have to be acquired at frequent intervals and on a year-round basis for this study. This can be done using the SAR aboard ERS-1. The European Space Agency Programme for International Polar Oceans Research (PIPOR) (ESA, 1985,1989) has requested SAR data in this area at three day repeat intervals during the 3-month commissioning phase of ERS-1, and at an interval yet to be determined (possibly 3 or 17 days) thereafter. The authors are co-PIs on this proposal and will receive the relevant data. Long term monitoring of iceberg flux through this 'choke point' could, therefore, be carried out through continuing acquisition and analysis of ERS-1 SAR data.

The authors of this report have the expertise in digital image analysis and interpretation necessary to undertake the long term monitoring of parameters 2 and 3 above and the links with the ERS-1 programme necessary for continuing data acquisition.

7. BIBLIOGRAPHY

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